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The Difference in Optics Between Volcanic and Plutonic Plagioclases, and its Bearing on the Granite Problem

By DORIS L. REYNOLDS

ABSTRACT

The orientation of the optical indicatrix in synthetic and volcanic plagioclases differs from that in plutonic and metamorphic plagioclases, the former " optics " being commonly known as " high-temperature ", and the latter as " low-temperature ". The difference between the two series is most marked at the sodic end and decreases towards An_{70}. The conclusion that the difference, though small, still persists in the range An_{70} to An_{100} is shown to be unwarranted on the evidence available. Grave doubt is thrown on the supposition that temperature is the controlling factor in determining the type of optics by Tuttle and Bowen's failure to synthesize low-temperature albite, even at low temperatures. It is known, however, that some low-temperature plagioclases (e.g. in albite-schists) have grown without the intervention of melt. The real antithesis may therefore be between plagioclases crystallized from melts, and those developed during the metamorphic reconstitution of rocks.

From the fact that the plagioclase, quartz, and alkali feldspar of undoubted magmatic rocks such as quartz-porphyry and rhyolite are respectively systematically different from those of granite it is concluded that the latter is a metamorphic rock.

I. THE EVIDENCE FOR TWO TYPES OF OPTICS

(*a*) *The range An_{35} to An_{70}.*

Eleven years ago Köhler reached the important conclusion that the orientation of the optical indicatrix in synthetic and volcanic plagioclases differs from that in plagioclases of plutonic and metamorphic rocks. This conclusion was the natural outcome of earlier work by other investigators showing that for plagioclases in volcanic rocks the poles of (010) form a scatter-band lying, for most of its extent, to the right-hand side of the relevant curve in Reinhard's (1931) Tafel 2. This curve of Reinhard's is constructed from univerasl stage measurements of plagioclases in plutonic and metamorphic rocks. For example, the scatter-bands formed by the poles of (010) plotted for plagioclases in Tertiary andesites from Roumania (Paliuc, 1932; Wenk, 1933) and North Borneo (Homma, 1932) lie well to the right of Reinhard's curve between the compositions An_{40} and An_{70}. Similarly for plagioclases ranging in composition from An_{41} to An_{100} in Pleistocene volcanic rocks from northern Patagonia, Larsson (1940, p. 364)

found that the poles of (010), although symmetrically dispersed in relation to Nikitin's (1933) curve down to an anorthite percentage of about An_{70}, fall more and more to the right of the curve with decreasing percentage of anorthite. In contrast Larsson shows (1) that the poles of (010) for plagioclases from gabbroic and noritic rocks of the Nygård pluton, Central Sweden, are symmetrically dispersed in relation to Nikitin's curve; and (2) that the scatter-band of poles of (010) for plagioclases from the hyperites and hyperite amphibolites from western Sweden is bisected by Nikitin's curve. Larsson thus clearly demonstrates, by means of the orientation of the optical indicatrix, a difference between the plagioclases from a suite of volcanic rocks on the one hand, and those from suites of plutonic and metamorphic rocks on the other.

In establishing the two types of optics Köhler adopted the highly sensitive method of determining plagioclases by measuring the angles between equivalent optical directions in related twin lamellae. The optical orientations of two related twin lamellae are plotted on a stereographic projection parallel to (010), following Reinhard's procedure; the angles $\alpha\alpha'$ and $\gamma\gamma'$ are then measured across (010), whilst the angle $\beta\beta'$ is the supplement of the appropriate angle measured across (010). These angles have become known as the Köhler angles. Additional angles between equivalent and inequivalent optic axes can also be measured, such as AA′, AB′, etc. The several measurements, if exact, should be mutually consistent and therefore reliable. By applying this method to the determination of plagioclases in volcanic rocks from Linosa and Estel, and constructing comparative curves from Reinhard's data, Köhler (1941) was able to demonstrate the difference in optical orientation between volcanic plagioclases on the one hand, and plutonic and metamorphic plagioclases on the other, for the composition range An_{35} to An_{54}. Köhler attributes the difference in optics to a difference in crystallization temperature, contrasting the known high temperature of crystallization of plagioclases in volcanic rocks with the inferred low temperature of crystallization of plagioclases in plutonic and metamorphic rocks. The two types of optics have consequently become known as high- and low-temperature optics.

In the absence of correlated chemical analyses and optical data for more calcic plagioclases from volcanic rocks, Köhler's co-workers, Tertsch (1941, 1942, 1943) and Scholler (1942), attempted to extend knowledge of the high-temperature series beyond Köhler's original restricted range in two ways: (1) by synthesizing plagioclases of the required compositions, and (2) by heating plutonic plagioclases of known composition; in both cases making the optical measurements necessary to extend the range of Köhler's curves. The second

method was suggested by Barth's (1931) discovery that there was a permanent change in the position and shape of the optical indicatrix of a plutonic labradorite, *ca.* An_{52}, after heating at 1000° C. for 300 hours. It is not the writer's purpose to present a historical account of these investigations, and the reader is referred for further details to the bibliography at the end of the paper, and particularly to the determinative curves for high and low temperature optics recorded by Köhler (1941) and Tertsch (1942).

In 1944 Oftedahl discovered that determination of plagioclase phenocrysts in the rhomb-porphyries of the Oslo region, by the zone method of Ebert, gave results from 7 to 9 per cent higher in anorthite than those obtained by the immersion method. Oftedahl finally explained this discrepancy by reference to Tertsch's comparative curves of the Köhler angles for high- and low-temperature optics. The Köhler angles of the rhomb-porphyry phenocrysts were found to agree best with the angles for high-temperature optics, thereby indicating a lower anorthite content more nearly in agreement with that determined by the immersion method. Subsequently Oftedahl (1948) established that high-temperature and transitional optics characterize the phenocrysts of the volcanic rocks of the Oslo area, and are also found in the plagioclases of some of the sub-volcanic rocks, and in those of intrusive akerite porphyries. He found the range of composition of the plagioclases with high-temperature optics in the volcanic rocks to be about An_{11} to An_{58}.

More recently van der Kaaden (1951) has made a comprehensive study of the optics of natural plagioclases, both volcanic and plutonic, from many localities. His data include 328 new determinations of optical orientations, together with a compilation of reliable measurements recorded by previous investigators. Van der Kaaden's plots of the poles of twin axes on stereographic projections perpendicular to β show : (1) that the poles of twin axes of plagioclases in volcanic rocks form a well-defined narrow band ; (2) that the poles of twin axes of plagioclases in plutonic rocks form a well-defined narrow band ; and (3) that the curves so formed, for volcanic and plutonic rocks respectively, are clearly separated within the composition range An_{35} to An_{70}, the curves for volcanic plagioclases lying to the right of those for plutonic plagioclases. Van der Kaaden's data include the measurements made by Köhler, but do not incorporate those of Schumann (1931), Paliuc (1932), Homma (1932), Wenk (1933), Larsson (1940), or Oftedahl (1944, 1948), presumably because these investigators provide no numerical record of the optical orientations of the plagioclases they investigated. Van der Kaaden's work therefore supplements that of these other investigators in demonstrating that the optical orientation of plagioclases in volcanic rocks differs from that

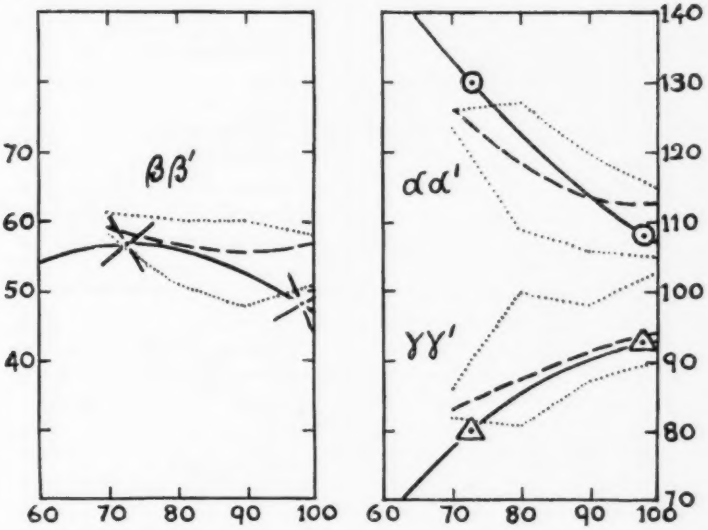
of the plutonic and metamorphic plagioclases represented by Reinhard's (1931) standard curves. Moreover, van der Kaaden confirms Reinhard's curves with 146 measurements of the orientation of the optical indicatrix in plutonic plagioclases, just as Larsson confirmed them with more than 300 measurements of the optical orientation in plagioclases from Swedish plutonic and metamorphic rocks. The combined work of all investigators thus provides a convincing demonstration that the orientation of the optical indicatrix in plagioclases from volcanic rocks differs from that in plagioclases from plutonic and metamorphic rocks for the range from An_{35} to An_{70}, the difference decreasing with increasing anorthite content.

(*b*) *The range An_{70} to An_{100}.*

Tertsch (1941, 1942) attempted to establish the Köhler angles and corresponding curves for high-temperature optics over the range An_{70} to An_{100} by determining the orientation of the optical indicatrix in synthetic plagioclases, crystallized from melts of composition An_{70}, An_{80}, An_{90}, and An_{100}, showing albite, carlsbad and albite-carlsbad twinning. He records twenty-eight determinations of Köhler angles and the average position of the optical indicatrix for An_{100} (1941), fourteen optical-orientation determinations of plagioclase crystallized from melt of composition An_{90}, ten of plagioclase from melt of composition An_{80}, and four of plagioclase from melt of composition An_{70} (1942, Tabelle 1, p. 199). From these determinations he calculated average values of the respective Köhler angles for supposed An_{70}, An_{80}, and An_{90} and for An_{100} (Tabelle 5, p. 204). Comparison of Tabelle 5 with Tabelle 1 (Tertsch, 1942), however, shows that the Köhler angles for albite twins of supposed composition An_{70}, An_{80}, and An_{90} derived from average positions of the respective indicatrices differ from those derived from the determined orientation of actual albite twins of supposedly the same compositions. For example, the values of $\alpha\alpha'$ derived from direct measurements of the optical orientations of plagioclases supposed to be An_{90} are 115, 116, and 117 (Tabelle 1), whereas the value recorded for $\alpha\alpha'$ (Tabelle 5) from an average position of the optical indicatrix is 113·3. These discrepancies led the writer to plot the Köhler angles for albite twins for each of Tertsch's measured orientations of the optical indicatrix for supposed An_{70}, An_{80}, An_{90}, and An_{100} in order to determine the limits of error within which his average values fall.

In Text-fig. 1 the average values for $\alpha\alpha'$, $\beta\beta'$, and $\gamma\gamma'$ (albite twins) determined by Tertsch from all his measurements of the optical orientation of supposed An_{70}, An_{80}, An_{90}, and An_{100} are plotted, and connected by dashed lines ; the dotted lines show the limits of error within which these average values fall. The firm lines of Text-fig. 1

connect the Köhler angles for An_{97} and An_{72} derived from Reinhard's data. The whole purpose of Tertsch's investigation was to draw a distinction, by means of the Köhler angles, between supposed high-temperature optics measured on synthetic plagioclases, and supposed low-temperature optics measured on plutonic and metamorphic plagioclases. Reference to Text-fig. 1 shows that no such distinction



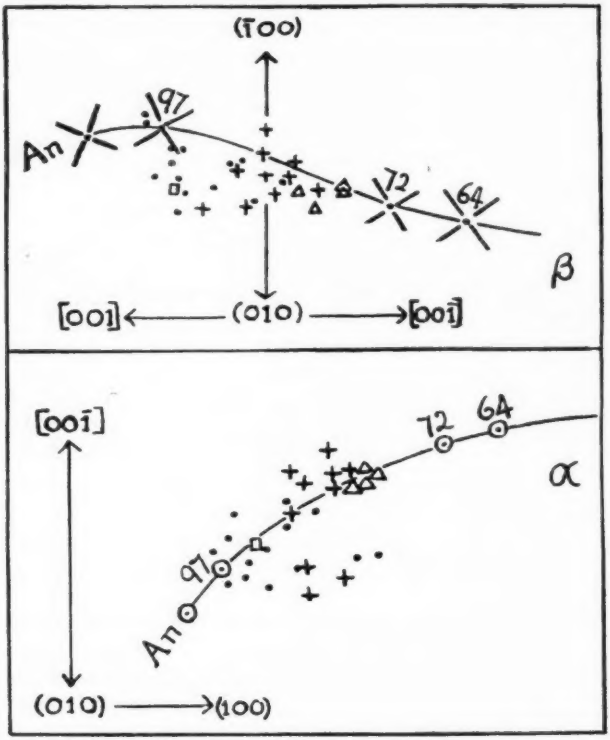
TEXT-FIG. 1.—Abscissae represent percentages of anorthite; ordinates represent Köhler angles.

The dashed lines are Tertsch's curves of Köhler angles for supposed high-temperature optics; they connect average values of these angles determined from synthetic plagioclases. The dotted lines indicate the range of the limits of error. The firm lines are the curves of Köhler angles, derived from Reinhard's data, for plutonic and metamorphic plagioclases of supposed low-temperature optics.

has been established, for the curves for the synthetic and the plutonic and metamorphic plagioclases both fall within Tertsch's limits of error.

If the Köhler angles are always subject to such wide limits of error then their value becomes questionable. It is obvious, however, that the discrepancies between the different measurements—apart from those relating to An_{100}—may depend on variation in the chemical compositions of the plagioclases measured (i.e. equilibrium may not have been maintained when the melts were crystallized) rather than on errors of measurement. In order to investigate this matter further

Tertsch's measured orientations of the optical ellipsoid for his various synthetic plagioclases have been plotted on a Wulff net of 40 cm. diameter, with the plane of the paper parallel to (010) as in Reinhard's (1931) Tafel 3. In Text-fig. 2 the scatter in the positions of α and β



TEXT-FIG. 2.—Parts of a stereographic projection illustrating the scatter in the positions of the poles of α and β for synthetic plagioclase, shown in relation to Reinhard's curves (1931, Tafel 3). Squares = An_{100}; dots = supposed An_{90}; crosses = supposed An_{80}; triangles = supposed An_{70}.

are shown relative to Reinhard's curves. For ease of reproduction the migration curves for α and β are shown with different orientations as indicated on the diagram. From Text-fig. 2 it is evident that for each supposed composition the orientations of α and β, relative to fixed crystallographic directions, show a marked scatter. It is also evident that Reinhard's curves for plutonic plagioclases lie within the

scatter belts. Since the scatter for supposed An_{80} lies, in the direction of composition change, within the limits of scatter of An_{90}, it is highly probable that the plagioclases that crystallized from melt of composition An_{90}, for example, varied in composition. If Reinhard's curve is used as a standard of reference then the measured plagioclases crystallized from melt of composition An_{90} range in composition from *ca.* An_{97} to *ca.* An_{80}, whilst the measured plagioclases crystallized from melt of composition An_{80} range in composition from *ca.* An_{90} to *ca.* An_{80}, and those measured from crystallized melt of composition An_{70} range from *ca.* An_{81} to *ca.* An_{79}. These ranges strongly suggest that there was lack of equilibrium when the various melts crystallized. Indeed reference to the thermal equilibrium diagram for the plagioclases (Bowen, 1913) shows that An_{97} is actually the first plagioclase that crystallizes from melt of composition An_{90}. This suggestion is confirmed by Tertsch's descriptions of the plagioclases crystallized from the various melts. Lack of zoning and absence of residual glass appear to have been taken as sufficient guarantee that the synthetic plagioclases had the same composition as the respective melts from which they crystallized, in spite of the fact that more than one generation of crystals was present, as indicated by small crystals filling the interspaces between larger ones, and the occurrence of skeletal crystals.

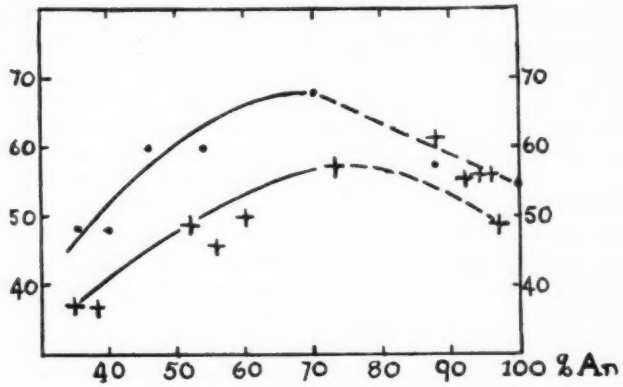
Whereas it can thus be concluded that Tertsch's (1942) investigation provides no evidence for the existence of two series of plagioclases, showing high- and low-temperature optics respectively, in the range An_{70} to An_{100}, the question still remains as to whether two series do in fact exist. In order to investigate this matter still further the Köhler angles for albite twins have been calculated from relevant natural plagioclases for which adequate optical data and chemical analyses are recorded ; the results are listed in Table I. In order to save space

TABLE I

Köhler angles for albite twins (additional to those recorded by Köhler)

Source of Plagioclase	Investigator	An %	$\alpha\alpha'$	$\beta\beta'$	$\gamma\gamma'$
Synthetic An_{100} .	Duparc and Gysin (1926).	100	113	53	91
Synthetic An_{100} .	Tertsch (1941) . .	100	113	57	94
Anorthosite . .	Nikitin (1933) .	96	113	56	92·5
Olivine-norite .	Miller (1935) . .	94	112	56	96
Olivine-gabbro .	Nikitin (1933) . .	92	113	55·5	92·5
Anorthosite . .	Lundegårdh (1941) .	88	115·2	61·5	96
Heckla lava . .	Sørensen (1950) .	88	118	57	87
Hypersthene-gabbro	Nikitin (1933) .	60	157	50	61
Anorthosite (?) .	Duparc and Gysin (1926).	56	151	46	57
Diorite . . .	Nikitin (1933) .	38·5	172	37	38

only the $\beta\beta'$ angles are plotted in Text-fig. 3 where they are shown in relation to Köhler's $\beta\beta'$ curves for (*a*) high-temperature optics in the range An_{35}, to An_{70}, and (*b*) low-temperature optics in the range An_{35} to An_{72}. The latter were derived by Köhler from Reinhard's (1931) data. The average value of $\beta\beta'$ for synthetic anorthite (An_{100}), derived by averaging the average values for this angle from data recorded by Duparc and Gysin (1926) and by Tertsch (1941), is plotted as an end-point for Köhler's $\beta\beta'$ curve for high-temperature optics. It may be noted in passing that the $\beta\beta'$ curve thus constructed for lime-rich



TEXT-FIG. 3.—Ordinates represent Köhler angles.
The upper firm line is Köhler's $\beta\beta'$ curve for high-temperature optics, and the lower firm line is his $\beta\beta'$ curve for low-temperature optics. The dots represent volcanic plagioclases, except the dot for An_{100} which represents an average value for synthetic anorthite. The crosses represent plutonic and metamorphic plagioclases (see text and Table I).

plagioclases lies well above the supposed high-temperature curve drawn by Tertsch from his measurements of synthetic bytownite and anorthite. The $\beta\beta'$ angle for anorthite from limestone blocks erupted from Vesuvius (Reinhard, 1931 ; Nikitin, 1933) is plotted as an end point of the curve for low-temperature optics. Apart from the Vesuvius plagioclase, all the $\beta\beta'$ angles for natural plutonic plagioclases plotted within the range An_{70} to An_{100} correspond best with the extension of the high-temperature curve. In other words the Köhler angles for synthetic An_{100} and volcanic An_{88}, lie within the limits of error of measurements on plutonic plagioclases. If there are two types of optics for lime-rich plagioclases it is thus doubtful whether it will be possible to discriminate between them by optical methods. Moreover, Tuttle and Bowen (1950, p. 578) record that natural plagioclases whose

compositions lie between An_{70} and An_{100} show only slight changes in optical and X-ray properties on heating.

(*c*) *The range An_0 to An_{35}.*

Eighteen years ago Beliankin (1934) investigated albite from the arch of a regenerator chamber used in glass-making. The crusts containing the albite were formed by reaction between the fire-clay bricks of the regenerator and the melt it contained. Beliankin found this albite to differ from normal plutonic albite in being optically negative with a small optic axial angle varying from 31° to 34° in different crystals. The refractive indices, $\alpha = 1{\cdot}525$, $\gamma = 1{\cdot}533$, were also found to be lower than those of plutonic albite. Beliankin explained the "anomalous" optical properties of the synthetic albite as due to "the particularly high temperature of its formation, which exceeds considerably the temperature of crystallization of normal albite-containing eruptive rocks". He continued: "This would suggest a comparison of albite, which has a small 2V, with sanidine, the difference being only that for the formation of the former still higher temperatures are necessary than for the formation of the latter."

Two years ago Tuttle and Bower (1950) confirmed these early observations of Beliankin by showing that synthetic albite differs from natural plutonic albite in being optically negative with a small optic axial angle of about 45°, and in having lower refractive indices, $\alpha = 1{\cdot}527$, $\gamma = 1{\cdot}534$. They also recorded that phenocrysts of sodic plagioclase from rhyolites are related to synthetic albite in their optical properties.

By means of X-ray powder photographs, Tuttle and Bowen showed that low-temperature plagioclases are easily distinguishable from the corresponding high temperature forms if the composition is near albite, but that it becomes increasingly difficult to establish a difference as the composition approaches An_{10}. The small optic axial angles of members of the high-temperature series of sodic plagioclases increase with addition of anorthite and become essentially the same as that of the low temperature series at An_{35}.

It is thus evident that there is a marked difference between high- and low-temperature plagioclases within the range An_0 to An_{35}, that high- and low-temperature forms can still be distinguished within the range An_{35} to An_{70}, but that it is not yet established whether or not this difference extends into the range An_{70} to An_{100}.

II. The Significance of the two Types of Optics in Relation to the Granite Problem

As Köhler (1949) has already pointed out, the fact that it is possible to distinguish between plagioclases of volcanic and plutonic origins

may be of great importance in unravelling the geological history of rocks whose origin is unknown. In particular it has a bearing on the origin of granite. A majority of petrologists—magmatists—regard granite on the one hand and quartz-porphyry and rhyolite on the other as products of crystallization of acid magma. Granite they regard either as a result of slow crystallization of acid magma in depth or as a consequence of granitization of country rocks by emanations from such magma. A minority of petrologists—transformists—whilst naturally agreeing that rhyolite and quartz-porphyry have crystallized from acid melts, regard granite (for reasons that need not be recapitulated here) as a product of metamorphism of pre-existing crustal rocks.[1] The argument between magmatists and transformists has much in common with the discussion as to which came first : the hen or the egg. Magmatists regard acid magma as the antecedent of granite, whereas transformists regard granite as a step towards the evolution of acid magma, the latter resulting from the fusion of granite. The discovery that the plagioclases of granite and metamorphic rocks differ from those of volcanic and hypabyssal rocks, and that the plagioclases of the latter rocks are identical with those synthesized from melts in the laboratory, should go a long way towards settling the disagreement.

When, in a recent investigation of high-temperature albite, Tuttle and Bowen (1950) discovered that synthetic albite has X-ray and optical properties different from those of natural albites from plutonic rocks, they attempted to synthesize the natural low-temperature variety. All their efforts met with failure ; even though they attempted the synthesis at a temperature as low as 250° C. under pressure of water vapour, the crystals that formed were still the so-called " high-temperature " variety. In these initial attempts to synthesize low-temperature albite Tuttle and Bowen crystallized glass of albite composition, but realizing that " glasses are predisposed to give high-temperature modifications, other initial materials, such as a mixture of quartz, corundum, and sodium silicate, were tried, but only the high-temperature form was obtained. Many runs were made, using various low-temperature albites as seeds, but no evidence of growth of low albite on the seeds was discovered. Again different initial materials were tried with the seeds, but with the same result " (Tuttle and Bowen, 1950, pp. 573–4). This failure of Tuttle and Bowen to synthesize low-temperature albite, even at low temperatures, throws grave doubt on the supposition that temperature of crystallization is

[1] It is fully realized that granites are commonly intrusive. Apart from granites in a volcanic setting, e.g. at Skaergaard (Wager and Deer, 1939) and Slieve Gullion (Reynolds, 1951), this is an expression of relative mobility in response to orogenic stress like that shown by intrusive limestones and pelites (Reynolds, 1949).

the decisive factor in determining whether high or low albite grows. It is true that natural low plagioclase can be made to invert to the high form by continuous heating (Barth, 1931), and the transitional optics recorded by Oftedahl (1948) may be explained in this way. Indeed, Tuttle and Bowen also found that natural low albite inverts to the high form during such treatment. They were unable to demonstrate experimentally, however, that the converse is true, i.e. that high albite will invert to low albite. It is therefore disappointing that, in spite of their own experience, Tuttle and Bowen should conclude an objective investigation by explaining the absence of high plagioclase from granite with such special pleading as " it is possible that the high form cannot survive the excessively long period of exposure to a moderately elevated temperature which is characteristic of plutonic conditions ".

The fact that metamorphic plagioclase is of the low type suggests that the fundamental difference between high and low plagioclase is that whereas the high variety crystallizes from a melt, low plagioclase grows by replacement in solid rocks. There is no possibility, for example, that the low albite of the albite-schists of south-west Scotland grew from a melt. From the fact that the puckered foliation planes of the schist are traceable through the low albites as trails of fine magnetite (cf. Clough, 1897, plate vi) it is apparent that the albites have grown metasomatically and replaced what was formerly schist. Examples such as this provide evidence against the assumption, made by Tuttle and Bowen, that low albite is a product of inversion of high albite ; the two varieties form under different geological conditions.

A further demonstration that low plagioclase forms otherwise than by inversion of high plagioclase is found in Sørensen's (1950) study of the plagioclase of some Heckla lavas. From 105 determinations of the optical orientations in plagioclases showing albite, carlsbad, albite-carlsbad, manebach, acline, ala, and pericline twinning Sørensen shows that both high and low optics are represented. Most of the lavas studied were fine-grained with small phenocrysts of plagioclase and aphanitic groundmasses of grain-size less than 0·01 mm. In these lavas the plagioclases have high optics. There were also a few coarser-grained rocks, however, in which the grain-size of the matrix exceeds 0·1 mm. From the presence of many vesicles in the relatively coarse-grained specimens, and the fact that the specimens from the oldest lavas were generally of the coarser-grained type, Sørensen concludes that coarsening of grain-size has resulted from devitrification of initially fine-grained lava.

In the devitrified lavas there are plagioclases showing (*a*) normal zoning in which a narrow outer zone has a slightly lower anorthite content than the core, and (*b*) inverse zoning, sometimes restricted to

one end of a crystal, in which an outer zone has a higher anorthite content than the core. The inverse zones have low optics whereas the cores on which they are grown have high optics like the plagioclases showing normal zoning. The unzoned plagioclases in these devitrified rocks, moreover, have the same composition as the inverse zones, and similarly exhibit low optics. The fact that the plagioclases showing low optics (An_{58}–An_{68}) are richer in lime than those showing high optics (An_{50}–An_{58}) demonstrates that the low optics have not resulted from inversion of high-optic plagioclase. It also strongly suggests that the devitrification process, responsible for the growth of the plagioclases with low optics, was a consequence of chemical changes involving enrichment of the lavas concerned in Ca and Al, probably at the expense of Na and Si (cf. Fenner, 1936).

Closely bound up with the difference in optics between volcanic and plutonic plagioclase is the recent discovery of Keith and Tuttle that rhyolitic and granitic quartz differ in the temperature of inversion from the α to the β form (Adams, 1950). In the words of the Director of the Geophysical Laboratory : " When classified according to the temperature of beginning of inversion, natural quartz samples fall into groups which are clearly related to the conditions of origin. For quartz phenocrysts of rhyolites and related intrusive porphyries and welded tuffs, the inversion temperatures are clustered around 572·7° C. Quartz from granites inverts at higher temperatures clustered around 573·2° C. Granite quartz and rhyolite quartz are completely separated on a plot of heating inversion versus cooling inversion " (Adams, 1950, p. 32).

In a recent paper full of mental somersaults, Tuttle (1952) discusses the contrasted mineralogy of acid volcanic and plutonic rocks. In addition to the difference in the high-low inversion temperatures of volcanic and plutonic quartz, Tuttle draws attention to other differences relating to types of inclusions and to the preferred orientation of quartz in many granitic rocks. He writes : " Preferred orientation of quartz is the rule in granitic rocks, although such an equant mineral would not be expected to be orientated by flow. The quartz of granites commonly has the same orientation pattern as the quartz in surrounding metamorphic rocks. . . . The quartzes of many granites, particularly the coarse-grained types, actually look like ' fragments ' of quartzite " (Tuttle, 1952, pp. 108–9). This is very good evidence for regarding granite as a metamorphic rock, but it is probably true only in relation to granites of a particular geological setting. It is, however, a shock to find that Tuttle, more influenced by his belief in granite magma than by the evidence he cites, concludes, without any supporting evidence whatsoever, that the quartz of plutonic rocks results from the recrystallization of quartz of volcanic type. The Director of the

Geophysical Laboratory, however, discusses the same problem with scientific objectiveness. With regard to the fact that granite quartz is separated from rhyolite quartz on a plot of heating inversion versus cooling inversion he writes (Adams, 1950, p. 32) : " These results can be reconciled with the magmatic theory of the origin of granites by assuming "—the important words are " reconciled " and " assuming " —one or other of the following : (1) that the quartz phenocrysts of rhyolite do not represent intratelluric crystallization ; (2) that a part, at least, of the quartz of granites actually grew at a lower temperature than the quartz phenocrysts of rhyolites ; and (3) that granites, which may initially have had some rhyolitic quartz, have all recrystallized at a lower temperature. The first of these assumptions he rules out because (*a*) many rhyolites represent explosive activity and there would be little possibility for phenocrysts to grow, except before eruption, and (*b*) the phenocrysts show evidence of dissolving, not growing, under the extrusive conditions last prevailing. The second assumpton he regards as equally untenable because, although the method employed can detect less than 5 per cent of one quartz mixed with another with different inversion characteristics, there is no evidence of more than one kind of quartz in the thirty investigated specimens of granite. Only the third *assumption* then remains as a plausible way of *reconciling* the facts with the magmatic hypothesis, i.e. it has to be *assumed* " that granites . . . have all recrystallized ". Now there is and can be no evidence for granite magma (as distinct from rhyolite magma) other than the existence and relationships of granite ; granite magma is and always has been a purely hypothetical concept. The discovery that the plagioclase and quartz of granite cannot be explained as direct crystallizations from magma is a death blow to the hypothesis.

Dr. Adams goes on to discuss the alternative possibility that granitic rocks are formed by a metasomatic process. For some reason, not stated, he assumes that two types of quartz would be present in granites if they result from the metasomatic alteration of pre-existing rocks. Perhaps this assumption depends on the fact that magmatists suppose that the small amount of granitization to which they subscribe is implemented through the agency of granite magma. The presence of only one type of quartz is certainly inconsistent with the supposition that granite magma is responsible for granitization unless Adams' third assumption, that there has been subsequent recryallization of the magmatic component, is still applied. The presence of only one type of quartz in granite, and that of a different type from rhyolite quartz is, however, consistent with and gives additional strength to the transformist interpretation of granite as a metamorphic rock evolved as a consequence of chemical interchanges (with relatively small additions and subtractions) within pre-existing geosynclinal rocks. If, as the

transformist interpretation implies, magma is in no way involved in the formation of granite, granite would not be expected to contain magmatic (rhyolitic) quartz.

Also of importance in this connection is the chemical difference between the alkali feldspars of volcanic and plutonic rocks. Tuttle (1952), by reference to histograms showing the relation between the number of chemical analyses and the composition of alkali feldspars in volcanic (including hypabyssal) and plutonic rocks, has recently demonstrated that whereas in volcanic rocks the alkali feldspars range in composition from Or_0 to Or_{75}, a majority being concentrated within the range Or_{40}–Or_{50}, in plutonic rocks they form two groups, separated by a hiatus, with the maximum concentration for potash feldspars near Or_{72} and for soda feldspars near Or_{12}. Experimental work by Bowen and Tuttle (1950) has demonstrated that at high temperatures, in the presence of water vapour under pressure, soda and potash feldspars, as already indicated in studies of dry melts, form a complete series of solid solutions of the type having a minimum-melting mixture. When crystallized at low temperatures, however, the complete solid-solution relation no longer obtains. In Bowen and Tuttle's own words, " unless some new factor intervenes—for example, critical phenomena —there will be a pressure of water vapour which will lower the solidus-liquidus curves to temperatures at which they will intersect the solvus. At this isobar and for all isobars corresponding to higher pressures, the feldspar diagram will cease to be one showing complete solid solution. Instead there will form two series of solid solutions with a hiatus."

Since the solvus has a maximum temperature at 660° C. $\pm$ 10°, it would be necessary for the minimum temperature of the liquidus-solidus curves to be lowered at least to this temperature before potash and soda feldspar could form, side by side, as direct crystallizations from magma. In this connection Perrin and Roubault (1951) in discussing the recent work on feldspars, have constructed a graph from Bowen and Tuttle's data showing the lowering of the minimum fusion point of the liquidus-solidus curves with increased pressure of water vapour. From this graph they conclude : " L'extrapolation ne permet pas de déterminer si la courbe tend vers une asymptote aux pressions extrêmement élevées ; mais on en déduit cependant aisément qu'à une pression de 4000 atmosphères correspondant à une profoundeur de plus de 14 km. le point de fusion minimum serait encore au moins de 650° C." As Perrin and Roubault indicate, it is not possible to extend the actual curve with exactness ; examination of Bowen and Tuttle's data shows, however, that log T varies in a linear way with log P, and from this relationship it appears that a water vapour pressure of 7,410 kg./cm² would be necessary in order to lower the minimum

point of the liquidus-solidus curves to the maximum temperature of the solvus. For Or_{75} and Or_{12} (shown by Tuttle to be the characteristic alkali feldspars of plutonic rocks) to crystalize directly from magma, a temperature as low as 550° C. is necessary, and to lower even the minimum point of the liquidus-solidus curves to this temperature a water vapour pressure of 35,480 kg/cm² would be necessary. This latter pressure corresponds to the weight of rock at a depth of over 60 miles. No wonder Tuttle (1952, p. 116) writes : " The probability of microcline and albite, of the composition found in many granites, crystallizing from a granite magma side by side is exceedingly remote. The association of relatively pure albite and microcline suggests temperatures so low that an origin other than magmatic is indicated unless exsolution can be shown to be operative."

Tuttle's appeal to exsolution is based on the experimental experience that if a synthetic feldspar, crystallized at 700° C. and higher, is " held at a temperature where it should form two feldspars, as indicated by the solvus, unmixing will occur in a period of a few days if water-vapor under pressure is present to facilitate mobility " (Bowen and Tuttle, 1950, p. 500). Moreover, " an intergrowth of two feldspars thus obtained by unmixing from a single feldspar can be re-homogenized by heating for about 24 hours at 700° even in the absence of water vapor." The presence of two feldspars in the product obtained by this induced unmixing cannot, like the intergrowths in natural perthite, be observed with the microscope ; they are revealed only by X-rays. Bowen and Tuttle go on to say : " We must now emphasize, however, that when examined further, the actual relations in the natural feldspars are considerably more complex than those in our synthetic analogues. These latter, throughout all the changes to which we have subjected them, retain their characteristics as high-temperature feldspars. . . . The natural intergrowths, on the other hand, are, in many instances, low-temperature feldspars. When these feldspars are heated . . . they give, within certain limits of composition, a homogeneous feldspar after a short period of heating, but the feldspar so produced is quite different from one grown at a high temperature or produced by re-homogenizing an intergrowth of two high feldspars. The optical relations of these natural feldspars heated only for the short period adequate for homogenization are those characteristic of low feldspar . . . Only after prolonged heating at temperatures but little below its melting temperature can a homogenized natural feldspar of the low-temperature variety be transformed into a feldspar having the optical properties of a high-temperature feldspar." It is therefore not a little alarming to find Tuttle so anxious to defend the magmatic origin of granite that he not only claims, without reservation, natural perthites as products of unmixing, but also regards the potash and soda feldspars,

occurring together in granitic rocks, as a result of unmixing of initially homogeneous feldspar.

The evidence is conclusive that the calc-alkali plagioclases, the alkali feldspars and the quartz of granite can no longer be regarded as primary crystallizations from magma. Petrologists who wish to conserve a magmatic origin for granite now have to abandon their previous arguments, based on textural relationships, that granite is evolved via a liquid line of descent from a more basic magma. As Tuttle says : " If the rock has undergone recrystallization of some components, inversion in all major phases, and unmixing in a major constituent, it is futile to draw detailed conclusions concerning an order of crystallization from the mutual grain relationships." They have, moreover, to agree with Tuttle " that granites and syenites containing relatively pure albite and orthoclase are actually metamorphic rocks which formerly contained only alkali feldspar ". Tuttle adds : " It is not certain whether the metamorphism responsible for the unmixing is the result of slow cooling (autometamorphism) or whether post intrusion regional and contact metamorphism have been operative."

Granite, according to the findings of the Geophysical Laboratory, is thus a metamorphic rock. From the evidence of its feldspars it falls into the same group as paragneisses and schists. There is already a wealth of evidence indicating that, like these latter rocks, it is evolved as a consequence of chemical interchanges (with relatively small additions and subtractions) within pre-existing geosynclinal rocks.[1] It is immaterial, in the present connection, whether these chemical changes are a consequence of solid diffusion, or of diffusion of ions through water or water vapour with which the rocks are saturated, or by some other process. The important consideration is that the mineralogical evidence is against the assumption that pre-existing acid magma was concerned in the formation of granite. Only if the temperature rose high enough to fuse the metamorphic granite would acid magma form, and this, after migration to higher levels in the earth's crust, would consolidate as quartz-porphyry or rhyolite characterized by minerals different from those of granite but resembling those synthesized in the laboratory from melts.

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The Ontogeny of the Trilobite *Dalmanitina olini*

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(PLATES IX AND X)

Abstract

The ontogeny of *Dalmanitina olini* Temple (Upper Ordovician) is described on the basis of specimens from Scania and North Wales. The development is similar to that of *D. socialis* (Barrande), but good preservation has enabled some detailed changes in the cephalon to be followed. The protaspid and early post-protaspid stages bear an anterior pit in each axial furrow opposite the frontal lobe. The protaspid anterior glabellar furrow becomes joined to this pit by an oblique furrow, and forms with the latter a composite oblique furrow which develops into the adult anterior glabellar furrow. This mode of development is considered probably to be common to all Phacopids : the significance of the anterior pits in this family is discussed. Preliminary observations are made on the growth-rates of various parts of the glabella, and the frontal lobe is found to grow in length (sag.) at a rate approximately equal to that of the rest of the glabella.

Introduction

DALMANITINA OLINI Temple (1952, p. 20) is a trilobite characteristic of the lower Brachiopod Beds (Upper Ordovician) of Scania, Southern Sweden, which has hitherto been considered as belonging to the doubtfully recognizable species *Phacops eucentra* Angelin, and which was described as such by Olin (1906) and Troedsson (1918). It is found also in Britain in the Ashgillian rocks of the Lake District (upper "*Phacops mucronatus*" Beds at Coniston ; Marr, 1916) and of North Wales (Deganwy Mudstones at Conway ; Elles, 1909), but until recently in this country it has not been distinguished from the closely related species *Dalmanitina mucronata* (Brongniart).

The ontogeny of *D. olini* was first described by Troedsson (1918, pp. 64–7, text-fig. 10). Some early growth-stages were discovered recently near Conway, and it was felt desirable to compare them with those from Scania and to supplement to some extent the earlier figures and description of the development. Through the courtesy of Professor Troedsson and the authorities of Lund Palaeontological Institution, the Swedish specimens have been placed at my disposal for study. My thanks are due to Professor Troedsson for permission to examine his collections, and to Dr. O. M. B. Bulman under whose supervision the work was carried out. Dr. Bulman and Dr. C. J. Stubblefield have read the manuscript and have offered much constructive criticism. I gratefully acknowledge receiving, during the period of the research, a grant from the Department of Scientific and Industrial Research.

Material

Immature forms of *D. olini* are known from Scania, Conway, and Ash Gill Beck near Coniston, and from them a complete series from protaspis to adult has been assembled. All these occurrences belong to the zone of *D. olini* (" sub-zone of *D. eucentrus* " ; Troedsson, 1918) and in each instance the fauna of the including bed consists almost entirely of *D. olini*. It has been assumed that the developmental series belongs to one species only, here considered as being *D. olini*, but a similar history might apply equally to *D. mucronata*, although undisputed young forms of the latter have not been found. The differences between the two species are relatively small and may be apparent only in fairly adult stages, while the development described below is concerned essentially with immature individuals.

The specimens from Conway were found in the Deganwy Mudstones, one mile south-west of Conway, in a steep lane running uphill from Hendre village towards Oakwood Park Hotel. Immediately south of Bryn-hendre, the mudstones outcrop on both sides of the lane, weathered to a greyish-brown on the south-west bank, but blue and hard near the wall to the north-east, where they yield many growth-stages of *D. olini* together with a few crinoid columnals and some ostracods. Adult *D. olini* from this locality have already been figured for comparison with Swedish forms (1952, pl. iii, figs. 6, 8). The trilobites are not greatly distorted, but their tests have weathered out and have been replaced by a yellow granular deposit, which does not retain well the fine detail of the smallest forms. For this reason the following descriptions are based mainly on the better preserved Scanian specimens collected by Professor Troedsson from the Brachiopod Beds of Röstånga and Tommarp (Troedsson, 1918, pp. 10, 17). The material from Ash Gill Beck is similar to that from Conway and Scania, but is mostly distorted, and has not been used in the descriptions. All the specimens are preserved either in the Palaeontological Institution, Lund (registration numbers prefixed by LO), or in the Sedgwick Museum, Cambridge (prefix SM).

The figures on Plates IX and X are reproductions of crayon drawings made on camera-lucida outlines ; it is possible to represent the fine detail of small specimens more clearly in this way than by means of direct photographs, as the great convexity of the early stages makes high-power photography very difficult.

Terminology

In the descriptions that follow, the glabellar furrows and lobes are numbered from back to front, excluding the occipital furrow and ring. In order to save repetition of the phrase " transitory pygidium " in describing immature forms, the pygidial shield at all stages of

development is called, for convenience, pygidium. The contractions (tr.), (sag.) and (exs.) are used to indicate that the measurements so qualified are respectively transverse to the axial line of the trilobite ; along the axial line ; and parallel to, but outside, the axial line (R. and E. Richter, 1940, p. 16). Where the lengths (sag.) of cephala are quoted they are glabellar lengths, exclusive of the anterior border : the stated lengths of protaspes do not include the pygidia, and are thus comparable with those of post-protaspid cephala. Pygidial lengths do not include the terminal spines.

Descriptions of Specimens

The smallest specimen is an internal mould of a protaspis, 0·51 mm. long, from Conway (SM : A 36169 ; Plate IX, fig. 1). It is badly preserved, but shows an almost parallel-sided axis with six annulations of which the last probably represents the pygidium. The cheeks are narrow in dorsal view, but when seen from behind they pass over a geniculation to form a vertical distal portion about twice as wide (tr.) as the inner part.

The next recognizable protaspid stage is represented by a number of specimens, including SM : A 36161 from Conway (Plate IX, fig. 2) and LO 61 JT from Scania (Plate IX, fig. 3 ; Plate X, fig. 6). This last is an external mould, 0·66 mm. long, but has been drawn as if it were the cast of that mould, while the description has been written as if the cast were being described. The specimen is well and delicately preserved, but has suffered some lateral compression.

The general outline is almost circular, and the whole shield is divided by two parallel transverse furrows, the posterior of which separates the pygidium from the cephalon. The central axis, representing the glabella and the axis of the pygidium, is appreciably raised above the level of the cheeks, but the axial furrows outlining it are not deeply incised.

The sides of the glabella are parallel between the furrow marking the posterior margin of the cephalon and the anterior pair of glabellar furrows, and over this region the glabella is a little less wide in dorsal view than each cheek. The frontal lobe is, however, half as wide again as the parallel-sided " stalk ". There are on each side of the glabella three furrows and an occipital furrow, the four furrows being equally spaced so that the occipital ring and the three posterior glabellar lobes are of the same length (exs.). The frontal lobe is about twice as long as each of these posterior lobes. The glabellar furrows extend on each side about a third of the way across the glabella. Each furrow is a shallow depression in the surface of the glabella, and is only slightly elongated transversely. The three posterior glabellar lobes are similar

in size and shape, although the first (posterior) shows a slight indication of being circumscribed by a very faint longitudinal groove running from the inner end of the first furrow to the occipital furrow.

The frontal lobe of the glabella is almost twice as wide (tr.) as it is long, being approximately oval in plan. It is bounded posteriorly by the third glabellar furrows and the axial furrows, and slopes quite steeply downwards in front. The actual anterior margin is partly broken, but there are traces of a horizontal border in front. From the inner part of each anterior glabellar furrow a faint longitudinal furrow runs forward over the frontal lobe in a slightly sinuous course to the anterior margin of the glabella. As a result there is cut off on each side of the frontal lobe a slightly inflated oval area, while the anterior margin of the frontal lobe is indented at the ends of the longitudinal furrows. A little behind the widest part of the frontal lobe each axial furrow is deepened to form an anterior pit. Between this pit and the anterior glabellar furrow the axial furrow is very shallow, so that the frontal lobe appears to be joined obliquely to the anterior part of the cheek.

The horizontal border in front of the glabella widens posteriorly towards the region of the anterior pits, behind which it stretches back on each side to a point approximately opposite the third glabellar lobes. This last part of the border is considered to be the protaspid palpebral lobe. It is separated from the cheek by a faint groove joining up with the anterior pit.

The occipital furrow has on each side a deep part similar to the glabellar furrows, and is continued across the glabella by a thin groove, which is convex forwards. The lateral parts of the occipital ring are more depressed than the glabellar lobes, but the axial portion of the ring is higher than the axis of the anterior part of the glabella. This elevated boss bears five tubercles, one central and the others at the corners of a square.

Each cheek consists of a nearly horizontal inner part, which passes distally over a fairly sharp geniculation into a vertical distal portion, although as the edges of the cheeks are broken, very little of the latter is visible. The inner parts of the cheeks are gently arched from side to side, and slope downwards anteriorly. They are defined posteriorly by the furrow marking the junction of the cephalon and pygidium. This furrow, which is parallel to the posterior marginal furrow, leaves the axial furrow at right angles, but bends back as it approaches the lateral margin. The posterior border of the cheek is thinner (exs.) than the occipital ring, as the posterior marginal furrow meets the axial furrow behind the junction of the latter with the occipital furrow. On the left side the posterior border is produced backwards into a short genal spine, which continues the curve of the lateral margin.

The inner parts of the cheeks show a pitted reticulate pattern affecting both surfaces of the test, but this is not found on the palpebral lobes, the posterior borders or the glabella. There is a single tubercle in the median line of each cheek just in front of the posterior marginal furrow, and several tubercles on the lateral parts of the frontal lobe, the palpebral lobes and the anterior margin.

The pygidium is semicircular in plan and is inclined downwards posteriorly, its posterior margin being incomplete. It is approximately twice as long (sag.) as each glabellar lobe, and is divided by the axial furrows into a central raised axis, the continuation of the glabella, and two lateral lobes corresponding to the cheeks. The axis is parallel-sided anteriorly, but its posterior part widens and is bulb-shaped. There appears to be an indentation on each side of the axis near its mid-length, and this delimits an anterior part about as long (sag.) as each glabellar lobe. The lateral lobes are less elevated than the axis and are arched down distally and posteriorly; they show on each side traces of a slight transverse furrow opposite the indentation of the axis. The pygidium is smooth.

Specimen LO 2914 t is an internal mould of a protaspis, 0·69 mm. long, which was figured by Troedsson (1918, pl. i, fig. 24), and is at about the same stage of development as that described above. The right side shows well the relation of the pygidium to the cephalon. The lateral margin of the pygidium meets the posterior margin of the cephalon at a point about two-thirds of the distance from the axial furrow to the genal spine, and both the distal part of the cheek and the genal spine are thus free of the pygidium. The axis of the pygidium shows two rings, and there is a short backwardly-directed spine halfway along each lateral margin. This spine belongs apparently to the anterior segment of the pygidium.

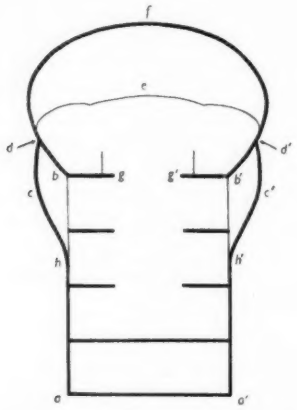
After the protaspis no complete specimens have been found, there being only isolated cephala and pygidia, which pass gradually by increasing size to the adult state. It is not possible to separate the meraspid and holaspid stages, nor to distinguish the degrees of the former; nor can the changes during development in the numbers of thoracic and pygidial segments be followed.

These post-protaspid stages, some of which are figured on Plates IX and X, do not show any features of outstanding interest that were not already present in the protaspis, and the gradual development of the protaspid features is described in the following summary.

Summary of Development

D. olini is closely related to *D. socialis* (Barrande), and as was

pointed out by Troedsson the main features of its ontogeny are similar to those of the Bohemian species described a century ago by Barrande (1852, p. 552*a*, pl. 26). The Scanian forms, however, are more finely preserved than the Bohemian, and show in detail the changes that occur during the early part of development. It is possible that some



TEXT-FIG. 1.—Superimposed diagrammatic outlines of the glabellas of a protaspis and a meraspis of *Dalmanitina olini* Temple from Scania, to show the relative changes in shape during early post-protaspid development.

Thin outline *ahbded'b'h'a'*—protaspis 0·66 mm. long (LO 61 JT).
Thick outline *ahcdfd'c'h'a'*—meraspis 1·60 mm. long (LO 4 JT).
The protaspis and meraspis are drawn to different longitudinal and transverse magnifications such that their respective glabellar stalks (*abb'a*) coincide. The widths of the stalks are drawn equal at a convenient value intermediate between those corresponding to the different magnifications of the two glabellas : the protaspis is thus too wide in proportion to its length, and the meraspis too narrow. The proportions of the various parts of each stage to each other in a longitudinal or transverse direction are correct, but all oblique proportions and all angles are distorted.
The points *d* and *d'* mark the positions of the anterior pits.

of these changes, particularly the evolution of the glabella, may apply also to the ontogeny of *D. socialis*.

Cephalon.

The changes in the form of the glabella are represented diagrammatically in Text-fig. 1. On this figure have been superimposed the glabellas of a protaspis 0·66 mm. long (*ahbded'b'h'a'* ; Plate IX, fig. 3) and a meraspis 1·60 mm. long (*ahcdfd'c'h'a'* ; Plate X, fig. 2).

The two glabellas have been drawn to different scales so that their glabellar stalks (*abb'a'*) may coincide : the figure shows therefore only the relative changes of form between the two growth-stages.

The most important change concerns the development of the adult anterior glabellar furrow, and has been completed in the meraspis of Text-fig. 1, which in this respect does not differ essentially from the adult. In the protaspis the anterior furrow *gb* is short, transverse and equivalent to each of the other glabellar furrows, with which there is no doubt that it is homologous. The axial furrow of the protaspis runs straight forward from the posterior margin of the glabella to the distal end of the anterior glabellar furrow (*a* to *b*) and then turns outwards to the anterior pit (*b* to *d*), this latter part of the furrow being very shallow. During post-protaspid development, the part *hb* of the axial furrow becomes very indistinct (Plate IX, figs. 4 and 5), while the third glabellar lobe merges with the adjacent part of the cheek and, since *db* is still very shallow, almost isolates the anterior furrow *gb*. At the same time the axial furrow is deflected around the enlarged third lobe, and runs into the anterior pit from a new direction *ahcd* (Plate IX, fig. 6 ; Plate X, figs. 1 and 2). The part *bd* of the protaspid axial furrow becomes the oblique outer part of the adult anterior glabellar furrow *gbd*, of which only the transverse inner part *gb* is equivalent to the anterior glabellar furrow of the protaspis. The anterior glabellar furrow of the adult is thus a composite feature of two-fold origin, and the anterior pit which in the protaspis is distinct from and in front of the anterior glabellar furrow is found in the adult at the distal end of that furrow. Although later changes in shape tend to straighten out the kink in the anterior glabellar furrow, the two parts of the furrow are usually distinguishable even in fully-grown forms of *D. olini* and *D. mucronata*, while the inner part *gb*, representing as it does the original glabellar furrow and corresponding to the other appendifer-bearing furrows, is often deeper than the outer *bd*. As another result of the deflection of the axial furrow around the enlarged third lobe, the second glabellar furrow, which in the protaspis reaches to the axial furrow, is in the adult either cut off completely from that furrow, or is joined to it only by a shallow groove. The first glabellar furrow is not affected by these changes.

The anterior pits, which are well developed in the protaspis and seem to be most marked in cephala about 1·5 mm. long, become shallower during later growth and do not occur in cephala longer than 2·8 mm., although their locus is marked in the adult by the junction of the anterior glabellar furrows with the axial furrows.

Text-fig. 1 also represents the great relative increase during growth in the length (sag.) of the frontal lobe of the glabella. Measurements show that the frontal lobe grows at a constant rate approximately four

times that of each of the other lobes, or at the same rate as the rest of the glabella taken together (the glabellar " stalk ") : so that between any two growth-stages the increase in length of the frontal lobe is equal to that of the stalk, or to half the total increase in glabellar length. There is, however, a constant very small difference (about 0·25 mm.) between the lengths of the frontal lobe and the glabellar stalk, the latter being the greater, but although this difference is appreciable in protaspid and early meraspid stages in which the stalk is relatively much longer than the frontal lobe, it is negligible in fully-grown cephala (Temple, 1952, pl. iii, figs. 5 and 6). If the growth-rate of the frontal lobe is constant for sizes smaller, as it is for those greater, than that of the smallest protaspis found (glabellar length 0·51 mm.), then the frontal lobe would be just as long as each glabellar lobe at a glabellar length of about 0·38 mm., and would disappear entirely at a glabellar length of 0·21 mm. : the latter on this assumption is a definite, the former a probable, limiting value for the total glabellar length of the smallest possible protaspis. While the anterior pits are present, the posterior part of the frontal lobe between them and the anterior glabellar furrows (the strip between the transverse lines *dd'* and *bgg'b'*) grows at a constant rate approximately equal to that of each of the glabellar lobes, while the anterior part grows at about three times this rate. The anterior pits maintain therefore during growth a constant position relative to the (protaspid) anterior glabellar furrows, and the great proportional increase in length of the frontal lobe is due entirely to that part of the lobe in front of the anterior pits.[1]

The twin longitudinal furrows on the frontal lobe of the protaspis disappear rapidly with increase in size, as do the indentations of the anterior margin : the frontal lobe of a cephalon 1·5 mm. long is unfurrowed, and its anterior margin smooth. A similar change affects the posterior margin of the occipital ring, which at one stage is quite angular (Plate X, fig. 2), but has become in the adult a smooth curve. The prominent central boss with five tubercles on the occipital ring of the protaspis [2] disappears later in *D. olini*, but it may be represented by the large tubercle on the occipital ring of adult *D. mucronata*.

The reticulation on the cheeks of the protaspis spreads rapidly to the central part of the glabella, the frontal lobe and the posterior marginal furrow, but not to the glabellar lobes, the occipital furrow nor to the posterior borders of the cheeks. Reticulation disappears in most cases during later growth, although it often remains on the

[1] It is hoped to present in detail the results of allometric studies on the development of *D. olini* in a subsequent publication.

[2] The arrangement of this group of occipital tubercles recalls the structure of the median " eye " in the centre of the glabella of *Tretaspis* which also is well marked in young stages, and for which a possible sensory function has been suggested (Størmer, 1944, p. 29).

cheeks as a coarse crenulation of both surfaces of the test even on the largest headshields.

In early meraspid stages, small hollow spines or tubercles, of which there are a few on the protaspis, begin to develop in abundance over the cephalon. These spines appear first on the lateral margins of the cheek where they point almost vertically, and on the anterior margin of the glabella where they point horizontally forwards. They spread quickly over the whole headshield, including the glabellar lobes which had not previously been ornamented, but do not occur in any of the furrows of the cephalon. The spines do not alter much in absolute size during later growth. They persist to the adult as pointed tubercles which become restricted to the outer surface of the test, the inner surface being smooth. In large forms the tubercles are most dense on the lateral and posterior borders and the doublures, and in some large individuals they disappear from other parts of the cephalon.

The protaspis is extremely convex from side to side, the distal parts of the cheeks being probably almost vertical, and the whole shield resembles a flattened hemisphere. During growth this convexity is gradually reduced as the cheeks lose their geniculation, and the adult cephalon is only slightly convex.

Pygidium.

The pygidium of the smallest, rather poorly preserved, protaspis (Plate IX, fig. 1) is very short and may have contained only one segment, but larger protaspes such as that of Plate IX, fig. 3 show two pygidial segments (making with the five segments of the glabella a total of seven for the whole shield). After this stage at which the pygidium is fused with the rest of the shield there is a gap, corresponding to early meraspid stages, in the pygidia represented in the collections, and the next undoubtedly complete forms are about 1 mm. long (Plate IX, figs. 4 and 5). In these pygidia, the pleurae are produced into long backwardly-directed spines which are the continuations of the posterior pleural bands.[1] These spines are hollow, and there is a doublure formed by the fusion of the inner ends of their ventral surfaces into a rod-like structure. The lateral lobes differ from those of the adult in that the oblique furrow of each pleura is very wide (exs.) and deep, and the anterior band very thin, although the posterior band is well defined. It is probable that a pygidium of this size was still transitory, and the anterior pleura of some specimens is apparently quite loosely attached

[1] This larval feature is retained to the adult in some Devonian Phacopids such as *Greenops* (Delo, 1940, p. 10). The formation of the spines from the posterior pleural bands may be contrasted with the condition in Paradoxidids and other Cambrian genera, where they form continuations of the anterior bands.

to that behind. In small pygidia of the order of 2 mm. long, the axis is already produced into a stout posterior spine.

With increasing size the free spines of the pleurae become smaller, although traces of them may still be seen in pygidia of *D. olini* up to 5 mm. long. The disappearance of the pleural spines is probably caused in part by resorbtion of their distal portions, and in part also by the distal widening of the doublure. The oblique furrows become narrower (exs.) as growth proceeds, while the anterior bands become better defined and comparable with the posterior.

The number of pygidial segments appears to become constant early in development, and after this there are no great departures from the adult number of seven.

Discussion

During the development of *D. olini* the loci of the anterior pits relative to the protaspid anterior glabellar furrows do not alter, and the pits remain segmentally in front of these furrows. The composite anterior glabellar furrows of the adult, the outer parts of which join the anterior pits to the protaspid anterior furrows, must therefore be oblique to the segmentation of the glabella. Although in this species they disappear during ontogeny, anterior pits remain to the adult in the phacopid genera *Chasmops* and *Pterygometopus*, where they lie at the distal ends of the adult anterior glabellar furrows, in the same positions as they are last seen in *D. olini*. This suggests that the oblique anterior glabellar furrows of *Chasmops* and *Pterygometopus* are homologous with those of *D. olini*, and have developed in the same way. The composite nature of the anterior glabellar furrow of *Chasmops* has previously been recognized by Öpik (1937, pp. 80, 81).

An oblique anterior glabellar furrow, similar to that of *D. olini* and *Chasmops*, is found in almost all Phacopids, and is probably of the same secondary origin throughout the family. In some forms, as in *D. olini* itself, the two parts of this furrow are distinguishable in the adult, there being often a change in direction between them, while in others this kink has been lost and the adult anterior furrow although oblique is quite straight. There are, however, some genera such as *Phacops* s. str. and *Trimerocephalus* in which the anterior furrow even of the adult consists of two distinct and quite unconnected parts, a condition which, in these otherwise advanced forms, may be neotenous. Such genera illustrate very well the origin of the complex phacopid anterior furrow, as may be seen by comparing for instance Barrande's pl. 22, fig. 34 with the meraspid outline of Text-fig. 1.

Anterior pits are found in many other families of trilobites including Bathyuriscidae, Illaenidae, Proetidae, Trinucleidae, Calymenidae, Cheiruridae and Encrinuridae. In the Cheirurids they are in the

axial furrows at the ends of the third glabellar furrows or even occasionally behind these furrows (*Reraspis* ; Öpik, 1937, p. 105), but in all other families where their relation to the segmentation of the glabella can be seen, the anterior pits lie in front of the third glabellar furrows and opposite the frontal lobe (or the fourth lobe) of the glabella. As this is also their original position in *D. olini*, and by inference in all Phacopids, anterior pits are positionally homologous throughout these trilobites. The oblique outer part of the anterior glabellar furrow of Phacopids linking the protaspid furrow to the anterior pit would then seem to be a feature distinguishing the Phacopids from all other trilobite families, with the possible exception of the Cheirurids. Ontogenetic studies of the latter family may help to determine whether the pits of cheirurid genera differ in their segmental position from those of other trilobites, or whether the anterior glabellar furrow of Cheirurids is of the same secondary origin as that of Phacopids.

The frontal lobe of the glabella of *D. olini* grows at such a rate relative to the other parts of the glabella that, in the adult of this species and also of *D. mucronata*, it is almost exactly as long (sag.) as the stalk of the glabella. This relation between the adult frontal lobe and the stalk is characteristic of typical members of the Phacopidae, and is found, more or less exactly and with few exceptions, throughout the family. It is not developed so constantly in any other group of trilobites, and in particular it distinguishes the Phacopids from the Cheirurids, in which the frontal lobe, although variously developed, is always shorter than the stalk. Some Encrinurids approach the phacopid condition, though not so consistently, while the lengthening of the encrinurid frontal lobe is not always accompanied, as it is in Phacopids, by a widening (tr.) relative to the stalk. Compared with the variability of these other proparian families, the constancy of the enlargement of the frontal lobe in the Phacopids is remarkable, and in conjunction with the peculiar nature of the phacopid anterior glabellar furrow emphasizes the homogeneity of the family.

If the ontogeny of *D. olini* reflects truly the phylogeny of the Phacopids, it is likely that the state of the protaspid glabella of this species, before the metamorphosis of the anterior glabellar furrows and the subsequent (and possibly consequent) loss of the anterior pits, represents a primitive ancestral condition. It is interesting to remark the resemblance of such a glabella to that of several adult Encrinurids, for instance *Cybele grewingki* Schmidt (Öpik, 1937, text-fig. 36), and indeed this glabellar plan is probably basic to all Encrinurids. The Phacopids may therefore have been derived, by specialization of the anterior part of the glabella and loss of the rostrum, either from early forms of the Encrinurids themselves or more directly from the common Cambrian ancestors of the Encrinurids and Cheirurids.

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EXPLANATION OF PLATES IX AND X

Growth-stages of *Dalmanitina olini* Temple from North Wales and Scania.

Note.—Spines and reticulation of the test have been omitted from all figures except Plate IX, figs. 3 and 4, and Plate X, fig. 6.

All specimens are internal moulds except where stated to the contrary.

PLATE IX

All figures × 22 approximately

FIG. 1.—Protaspis (SM : A 36169). Deganwy Mudstones, near Bryn-hendre, Conway.

FIG. 2.—Protaspis (SM : A 36161a). Same horizon and locality.

FIG. 3.—Protaspis—external mould (LO 61 JT). Brachiopod Beds, Röstånga (loc. 9, bed 2 : Troedsson, 1918, p. 12).

FIG. 4.—Meraspis (LO 64 JT). Same horizon and locality.

FIG. 5.—Meraspis (LO 2917 t). Same horizon and locality. The original of Troedsson, 1918, text-fig. 10 : 3.

FIG. 6.—Meraspis (LO 13 JT). Same horizon and locality.

PLATE X

FIG. 1.—Same specimen as Plate IX, fig. 6, × 11.

FIG. 2.—Meraspis (LO 4 JT), × 11. Same horizon and locality.

FIG. 3.—Meraspis (SM : A 36205a), × 11. Deganwy Mudstones, near Bryn-hendre.

FIG. 4.—Pygidium (LO 69 JT), probably of *D. olini*, × 15. Brachiopod Beds, Tommarp.

FIG. 5.—Pygidium—external mould (SM : A 36275b), × 15. Deganwy Mudstones, near Bryn-hendre.

FIG. 6.—Same specimen as Plate IX, fig. 3, × 45.

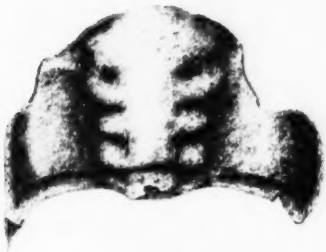
1

2

3

4

5



6

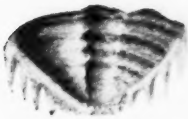
GROWTH-STAGES OF *DALMANITINA OLINI*.
(All × 22)

1 ×11

2 ×11

3 ×11

4 ×15



5 ×15

6 ×45

GROWTH-STAGES OF *DALMANITINA OLINI*.

A Tectonic Analysis of the Muness Phyllite Block of Unst and Uyea, Shetland

By D. Flinn

Abstract

The history of this block of low-grade metamorphic sediments begins with sedimentation, followed by strong deformation. Intrusion of spessartite dykes then took place, after which movement was resumed with deformation of the spessartites. Next followed jointing and the production of peculiar structures here called joint-drags. The final episode was one of block faulting, to which the formation of the Muness Phyllite Block itself as a graben is due.

Introduction

THE geology of Unst, the most northerly of the Shetland Islands, was described by H. H. Read (1930 and 1934), who showed that the island can be divided into a number of tectonic blocks of differing metamorphic histories. The blocks are separated by shear-belts along which a strong dislocation-metamorphism is produced. One of these blocks, called by Read the Muness Phyllite Block, forms the south-eastern part of Unst and extends into the island of Uyea (J. Phemister, 1930), which lies immediately to the south. The Block is bounded on the east by gabbro, pyroxenite, and serpentine of the Muness Greenstone and Serpentine Block and on the west by the gabbro of the Main Serpentine and Greenstone Block (Text-fig. 1).

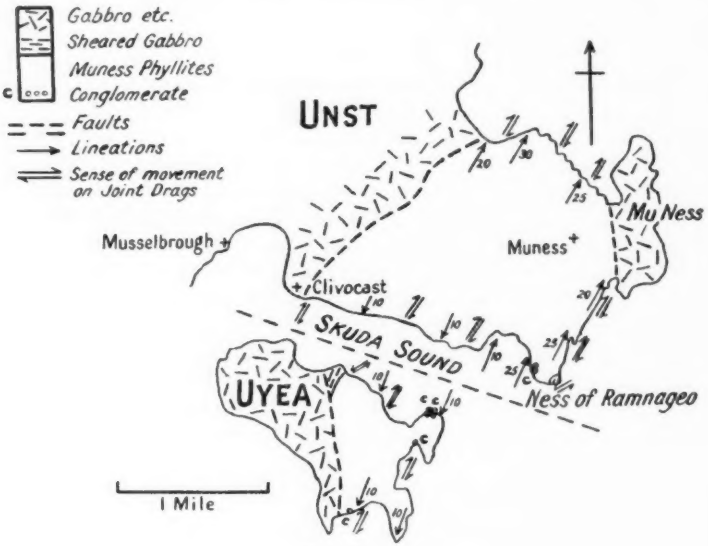
Coastal sections are very nearly the only exposures available; they present continuous exposures on the north and south edges of the Block in Unst and there are very good exposures along the shores of Uyea.

Lithological Composition of the Block

As the name given to it by Read suggests, the dominant rock-type of the Block is a *phyllite* which weathers rather easily and gives a first impression of uniformity of lithology within the Block. Closer inspection shows, however, that there are a number of distinguishable lithological types present. Most of these are characterized by their colours, e.g. grey-green with white stripes, black with brown stripes, dark green with brown stripes, and grey, that are a reflection of the mineral content; grey is due to sericite, green to chlorite, black to minute black particles visible in the thin-section, and brown to staining by an intergranular film of pigment. The striping is very fine, of the order of a dozen to the inch, and is perfectly parallel and continuous. It is parallel to large-scale lithological variations and can therefore be regarded as a sedimentary feature.

A less abundant *semipelitic rock* can be recognized by its greater

resistance to weathering and its more competent behaviour during deformation. Five exposures of *conglomerate beds* occur within the Block ; one is at Ness of Ramnageo in Unst (Read, 1934, p. 675), and four are on Uyea (Text-fig. 1). At Ness of Ramnageo, the conglomerate is 100 feet thick, those in Uyea are much thinner, the westernmost occurrence on that island being only a few feet thick. Among the contained pebbles are granitic and quartzitic rocks, pale greenish rocks, and speckled basic schists, lying in a greenish phyllitic matrix. Most of the pebbles are strongly elongated as described later. In places the conglomerate contains large numbers of small angular



TEXT-FIG. 1.

fragments that are seemingly undeformed, but these could be disrupted larger brittle pebbles.

At the south-east tip of Ness of Ramnageo is a small boss of *sheared gabbro* (Read, 1934, p. 676), very similar to the sheared rocks to be seen occasionally in the gabbro on either side of the Block. The contact between the phyllites and this gabbro is in some places sharp and in others gradational due to shearing. A spessartite intrusion, similar to those mentioned immediately, cuts across the contact.

Numerous *spessartite* intrusions (Read, 1934, p. 676) cut the Muness Phyllites and the adjacent gabbros. These form dykes a foot or two thick, which are approximately vertical and strike parallel to the

lineations of the phyllites. Most of them are parallel-sided continuous bodies, but some are observed to pinch out or to be openly folded. They weather easily, due to the mineralogical reconstitution and shearing which they have suffered.

Structural Elements

The phyllites show two well-marked surfaces of discontinuity, *S-surfaces*. The more common of these (S_1) is almost invariably parallel to, or separating, lithological variations and the colour striping, and can therefore be regarded as an original sedimentary feature. A schistosity due to parallel mica plates coincides with these sedimentary planes. In the field it is often difficult to recognize this bedding surface as it becomes obscured by the development of *folds* of all wavelengths from half a dozen to the inch to regionally-sized ones. A further complication is sometimes introduced by the presence of the second surface (S_2), a strain-slip cleavage, which itself is often gently folded. Both S-surfaces often contain lenticular quartz-segregations.

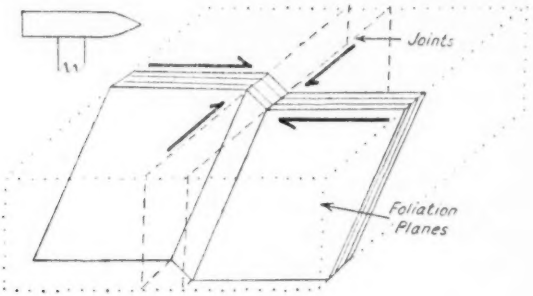
Much more obvious than the structures already mentioned are *lineations* which are the most noticeable structural feature of the phyllites (Read, 1934, p. 675).

These lineations are due mainly to the very small-scale folding of S_1 which has given so strong an anisotropism to the rocks that they fracture almost invariably parallel to the axes of folding. Thus many of the rock surfaces seen in the field show colour striping, schistosity, and quartz segregations parallel to one another and to the direction of the fold axes. Similarly the greatest dimension of fragments obtained by hammering records the same direction.

The direction of greatest *elongation of pebbles* in the conglomerate at Ness of Ramnageo averaged 025°, with a plunge of 24° towards the north. The pebbles are therefore elongated in a direction parallel to the axes of folding of the surrounding rocks (Text-fig. 1). More than fifty pebbles were measured. They showed rather considerable variation in their dimensions, probably due to varying resistance to deformation. The average proportions are 15·3 : 2·3 : 1. The pebbles were orientated so that they appeared to be flattened in the S_1 plane, with their smallest dimension perpendicular to that plane, and are elongated parallel to the lineation. It was therefore assumed that the pebbles were elongated by rolling, and that the tectonic directions of the above three proportions are $b : a : c$ respectively. A sphere of the same volume as the average pebble has a diameter of 3·3. With reference to such a sphere the pebble has been elongated 4·6 times in b, and shortened 0·7 and 0·3 times in a and c respectively.

Joints are ubiquitous and stand nearly always perpendicular to the lineation. They are locally in pairs, about an inch apart, with the

phyllites between them showing drag, as if movement on the two joints had been in parallel but opposite directions—as on the two sides of a fault (Read, 1934, p. 675). Similar structures in which the phyllites are as sharply and as regularly kinked in the same orientation and sense can also be observed (Text-fig. 2) ; these, however, are not accompanied by recognizable jointing. The axes of these monoclinal folds are entirely controlled by the orientation of the joints and of the



TEXT-FIG. 2.

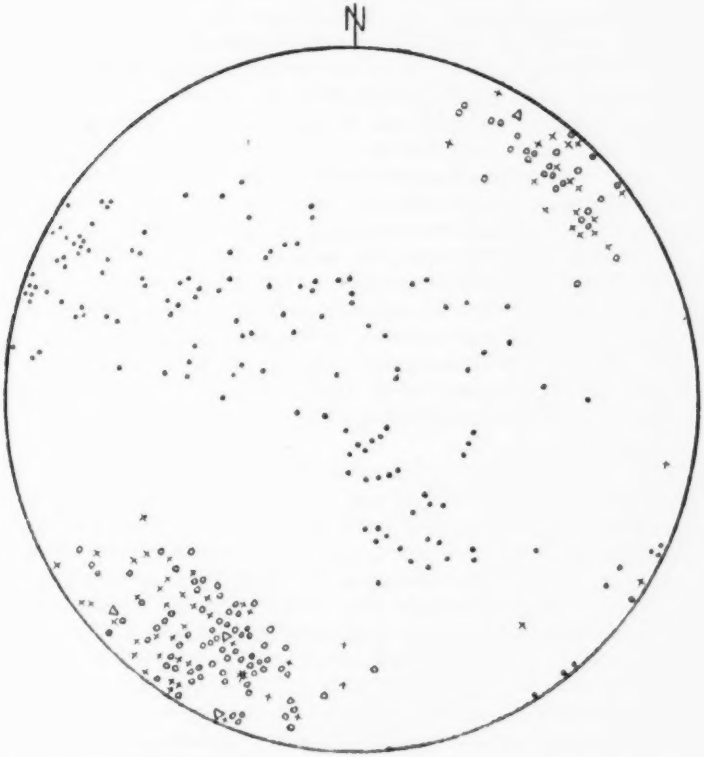
schistosity or bedding. These structures are here called *joint-drags*. Many other fractures also occur in the phyllites, but few show slickensides or other evidence of sense or direction of movement.

TECTONIC ANALYSIS

As a first step in the analysis, all structural observations were plotted on stereographic nets. Poles of those S-surfaces which are believed to be bedding or parallel thereto gave a slightly distorted girdle. The joints and lineations do not contribute to the girdle but, together with the elongations of the pebbles, form a maximum normal to it (Text-fig. 3). The gap on the east side of the girdle and the rather wide scatter of the joint-lineation maximum are due to the presence of an axial culmination in the Block. South of Skuda Sound the lineation plunges 10° southward ; north from the Sound the plunge changes from 10° southward to horizontal and then to 30° northward. The stereographic plot approximates to the theoretical form to be expected from a simply folded, originally planar system of beds, the lineations everywhere coinciding with the axes of the folds. The plot throws no light on whether the axial culmination is due to divergence of movement or to later superposed movements.

The simple movement picture illustrated in Text-fig. 3 shows that the Muness Phyllites were not subjected to any deformation earlier than that responsible for the major structural pattern. Throughout

the period of strong deformation the direction of movement was constant at any one place, but as shown by the axial culmination the direction may have varied slightly from place to place. The lineation everywhere forms the *b*-direction of this movement.



TEXT-FIG. 3.—Equal-area projection plot, upper hemisphere.

The position of the spessartites in the geological history of the Block may now be considered. Their injection can be regarded as a single act, as they are all identical in lithological appearance and in shape, attitude, and structure. They show, however, a variation in the degree of schistosity. They cannot have been injected before or during the main part of the period of strong deformation, because

they cut the structures caused by it and are, as compared with the phyllites, very much less deformed. They fill fractures that are mostly parallel to *bc*. Their direction is therefore unassociated with the direction of tension developed during folding and there appears to be complete independence between the main deformation and the emplacement of these intrusions.

The consolidated spessartites have, however, been subjected to movements which were weaker and more variable than the main movements. The observed effects are of four types ; mineralogical change, bending, fracturing, and the production of schistosity. The mineral changes are universal, strong, and complete, and result in the production of low-grade metamorphic assemblages. The bending or slight folding seen in some intrusions might be due to intrusion along curved fractures or to subsequent folding ; no jointing, thickening, or thinning of the intrusions was observed at these folds but a schistosity parallel to the axial plane of a fold is to be found in the intrusion which cuts the gabbro-phyllite boundary at Ness of Ramnageo. Spessartites are cut by and slightly offset by some joints and also by shears. The schistosity of the intrusions is parallel to their edges, so that in places the position of the margin is obscured. Schistosity across a dyke and parallel to a shear was observed in one locality.

Joints are a prominent feature of the Block. They are almost invariably *ac* joints. In that they cut the spessartites they belong to a later episode in the tectonic history of the Muness Phyllite Block. Two other facts support this conclusion. Distributed through the Block are a very large number of small shears which show movement but which rarely show evidence of offset. If all these fractures are plotted on a stereographic net it is seen that their poles form a dispersed girdle, occupying all the net except those two areas which would be occupied by the joint-poles if they were plotted on the same net. Although occasional offsetting of the dykes on a joint was noted, no shear zones were observed with an orientation parallel to the jointing. If joints had been present, some would have been active during this shearing phase. That no shearing took place on planes parallel to the *ac* joints seems to indicate that the rocks had not been weakened in that direction by jointing when the movements giving rise to the shearing were active. The intrusion of the spessartites and the production of the joints were therefore separated by a period of deformation by shearing.

The movement direction of the later deformation episode probably diverged little from that of the earlier, as is indicated by the following considerations. The joints, when they were developed later, formed perpendicular to the lineation which had been produced in the early deformation, despite the lapse of time and the resumption of

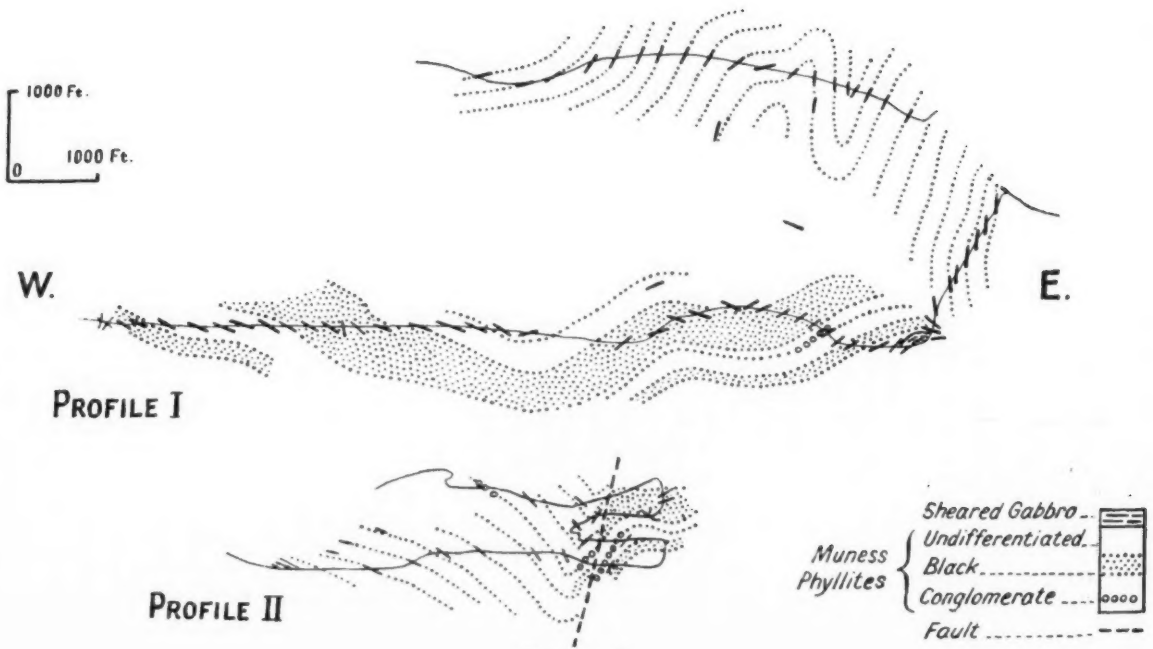
movement. The poles of the shears form on the stereographic net a dispersed girdle with a similar orientation to the S-surface girdle.

The joint-drags (Text-fig. 2) show within themselves two directions of movement. There is movement on the joint pair, in much the same manner as on a fault. There is also movement parallel to the strike of the S-surface; shortening has taken place in this direction. It is common for the offset between the joint pairs to continue for as much as ten yards along the strike of the joints. Joint-drags are frequently repeated one beside the other without a pause for several tens of yards. In no case was any slip observed to have taken place on either of the joints, in fact in the great majority of these structures no actual fracture was formed at all. Therefore the movement producing shortening parallel to the strike of the S-surface is considered to be of greater importance. A single joint-drag might be regarded as an incipient fault, but many hundreds occurring side by side over hundreds of square feet cannot be interpreted in this way. They must be the expression of a movement which can have a component parallel to the strike of the S-surface. No movement direction parallel to the *ac* joints—the movement which produced the early folding—could have a component in this direction. This movement therefore differed in direction from that causing the main deformation and probably from that giving rise to the later shear deformations.

The later deformation episode seems to be made up as follows. First, there was a reinstitution of movement, probably in the same sense and direction as during the earlier deformation episode but considerably reduced in effect. This movement gave rise chiefly to shears and to relatively slight deformation of the spessartites. It was followed by a movement of different direction which left only the widespread but insignificant joint-drags as evidence of its existence. At some time during the changeover the jointing took place.

A further advance in the tectonic analysis can be made by an examination of the actual structures rather than of their inter-relations. A section was drawn perpendicular to the axial direction in the manner described by C. E. Wegmann (1929, p. 107). Allowance was made for the arching of the axial direction. Separate profiles were drawn for Unst (Profile I, Text-fig. 4), and for Uyea (Profile II, Text-fig. 4). Lithology and structure although plotted separately, were found to agree rather well so that the profile was assumed to have some degree of validity.

Two features of the profile are immediately noticeable. They are the incongruity of structure and the differences in lithology between the northern and the southern portions of Profile I. The section along the south coast of Unst is well substantiated and it seems impossible to extrapolate this structure upwards in order to tie in with the north


1000 Ft.
0
1000 Fe.
W.
E.
PROFILE I
PROFILE II
Sheared Gabbro
Muness Phyllites
Undifferentiated
Black
Conglomerate
Fault

TEXT-FIG. 4.

coast structure without a complete change in style. No gradual change in the deformation was observed along the strike of the lineation ; the change in style takes place suddenly just north of the gabbro boss of Ness of Ramnageo. It may be suggested that a plane of discontinuity separates the two areas. The structure of the upper part is not inconsistent with this part having been thrust from the east to the west over the lower. It would be expected that the discontinuity would have been formed during one of the movement periods, most probably during the early one.

Examination of the gabbro-phyllite contacts provides material of interest, especially with regard to the nature of the dislocations bounding the Block. Gabbro-phyllite contacts are exposed at six points and can be grouped into two types. The first type is that of Ness of Ramnageo and that on the north coast of Uyea. At this latter contact the phyllites are nearly horizontal and rest on intensely sheared gabbro or chlorite-schist. This alteration of the gabbro continues for several hundred yards from the contact until good blocky-jointed gabbro is found. At Ness of Ramnageo no true gabbro occurs ; near the phyllites, which dip at 20° or less, is a rich green chlorite-schist, and elsewhere the gabbro is a sheared green and white speckled rock. At many points on the Ness it is impossible to fix the separation plane more accurately than within several inches. A number of powerful shears, one of which contains north–south slickensides, are tangential to the contact and cut both the gabbro and the phyllites.

The second type of gabbro-phyllite contact may be seen at the main boundaries of the Block in Unst. The dip of the phyllites varies between 30° and 90° and massively jointed gabbro is seen a few tens of feet from the phyllites. The gabbro at the contact is not schistose but sheared to a powder. In contrast to the first type of contact which forms headlands, this second group occupies small bays.

An interpretation of these two types of contact may be suggested as follows. At the first type, Ness of Ramnageo and the north coast of Uyea, the base on which the phyllites rested during much of their deformation is exposed—the tectonic contact may be a low-angle thrust. Both it and the associated fractures have suffered only slight movements since the earlier episode. It may be noted that the sense of the movements shown by the joint-drags is reversed at these first types of contact (Text-fig. 1). The second type of phyllite-gabbro contact, the main boundaries in Unst, and possibly on the south coast of Uyea, can be interpreted as normal faults throwing down the graben block of the Muness Phyllites between them. The map of Text-fig. 1 brings into prominence the cross-cutting nature of these two boundary-dislocations in relation to the axial direction. Since they cut obliquely

across all other structures, their formation can be considered the final episode in the tectonic history of the Block.

The island of Uyea in its present position would not fit structurally or lithologically on to the Profile I of Text-fig. 4. A good measure of lithological and some structural agreement can, however, be obtained by projecting it on to a place lower down so that the axial plane of its syncline coincides with that of the Muness syncline and the conglomerate horizon lies in its correct stratigraphical position. The absence of the conglomerate on the west limb of the Muness syncline may possibly be due to its wedging out to the west. It can therefore be proposed that a fault separates Unst and Uyea along the Skuda Sound (Text-fig. 1). This fault seems to throw down to the north and to have some rotation but little horizontal movement. It possibly causes the double 90° turn in the outcrop of the pyroxenite-serpentine boundary seen on Unst to the west of Musselbrough (Read, 1934, map, p. 663).

Tectonic Summary

The following is a summary of the tectonic history outlined above :—

1. Sedimentation.
2. Strong movement, giving the major part of the deformation, shearing, the tectonic contact between the phyllites and the underlying gabbro, and the discontinuity between the north and the south coasts of the Block in Unst.
3. Intrusion of the spessartites.
4. (*a*) Resumed movements probably in the same sense and direction as in 2, giving shearing, and the deformation of the spessartites.
   (*b*) Jointing.
   (*c*) Movement in a new direction, giving joint-drags.
5. Block faulting.

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The Lochaber Series South of Loch Leven, Argyllshire

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Abstract

Detailed mapping of a small area has shown that the Eilde Schist undergoes marked variations of facies when traced southwards from Loch Leven. Through the appearance of interbedded quartzite the normal schist of the western outcrop passes into a banded facies, while much of the banded facies of the eastern outcrop is believed to pass into massive quartzite mapped as Eilde Quartzite. This sedimentary transition, rather than tectonic causes, probably accounts for the rapid variations in thickness of the two stratigraphical units. An additional factor controlling the thickness of the Eilde Schist belonging to the eastern outcrop is the existence of an erosional surface at the base of the succeeding Binnenian Quartzite. The possible significance of these factors in the geology of the country north of Loch Leven is discussed briefly with special reference to some of the minor slides.

Introduction

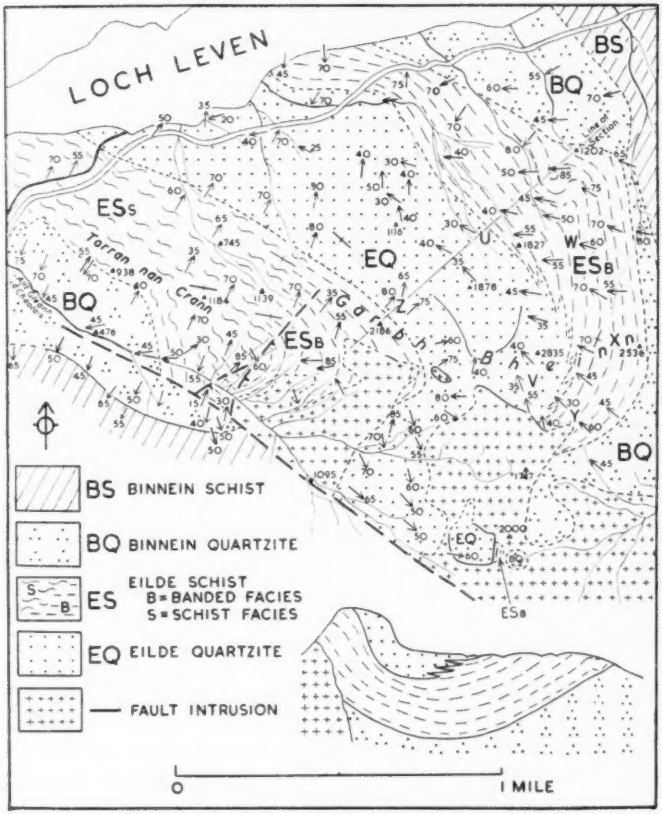
THE country around the eastern end of Loch Leven, which lies within the boundaries of Sheet 53 of the 1 in. Geological Survey map of Scotland, consists of alternations of quartzites and schists belonging to the Lochaber Series. This series is probably best regarded as belonging to the Moinian rather than partly to this group and partly to the Dalradian (Anderson, 1948, pp. 94–5).

The present paper, however, only deals with approximately three square miles of this area, centred on Garbh Bheinn (2,835 feet) on the south-eastern side of Loch Leven (Text-fig. 1). The Eilde Quartzite is the oldest member of the series to which reference is made, and this is followed in order of decreasing age by the Eilde Schist, the Binnein Quartzite, and the Binnein Schist respectively.

Structure.—The structure of the area has been fairly fully described by Sir Edward Bailey (1910, 1934). Briefly expressed, it can be said that the greater part of Garbh Bheinn consists of a very obvious asymmetric downfold, which pitches somewhat steeply towards the north-west. The core of this downfold is occupied by inverted Eilde Quartzite, which is clearly much thicker in the steep south-western limb than in the more gently inclined north-eastern limb. The quartzite is underlain by Eilde Schist and Binnein Quartzite (youngest), which outcrop in this order on the east and west flanks of the downfold. This inverted sequence of beds is proved by the attitude of the current-bedding present (Bailey, 1934, p. 485).

Immediately south of the crest of Garbh Bheinn the open fold structure rapidly closes to form an isoclinal fold with a south-westerly trending axis. The details of this structural change are shown in the outcrop map of Text-fig 2.

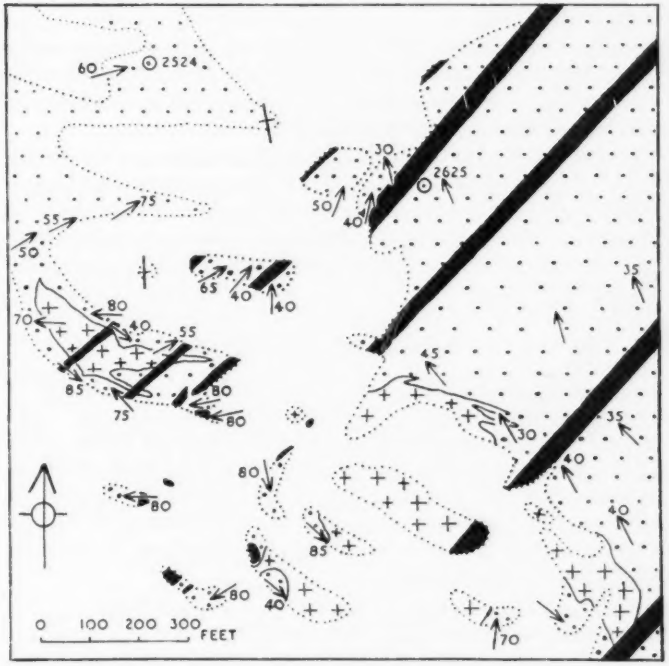
Although Bailey has never actually written about this structural change at any time it is quite clear, both from plate xlii of his 1910 paper as well as from his later maps and his unrecorded statements, that he has for long been aware of its presence. A glance at any of the



TEXT-FIG. 1.—Map of the area around Garbh Bheinn (dykes omitted). The section is drawn true to scale.

published maps should leave little doubt that this structural change must also coincide with a pitch culmination, otherwise the two outcrops of Eilde Schist would show signs of converging in the valley south of Garbh Bheinn. No such convergence is seen and, with a localized exception, the strike of the Eilde Quartzite on either side of the assumed position of the fold axis also remains fairly constant.

Towards the south the beds are cut across by two separate igneous masses belonging to the Lower Old Red Sandstone Glencoe fault intrusion. In the south-west the area is bounded by a powerful north-westerly trending fault, which Bailey (1934, p. 486) regards as being " in all probability, a branch of the boundary fault of the Glencoe cauldron subsidence ".



TEXT-FIG. 2.—Outcrop map showing the structural details of the Eilde Quartzite on the steep southern face of Garbh Bheinn. The fault intrusion is represented by crosses and the dykes are shown in full black.

Mapping.—The area has been mapped on a scale of 12 inches to the mile with parts on still larger scales. The four chief modifications to the existing maps are as follows.

(1) The small outcrop of Moines in the south-east of the area, which is surrounded by the fault intrusion, is now regarded as consisting of Eilde Quartzite as well as Eilde Schist. Formerly the whole of the outcrop was mapped as Eilde Schist (Bailey, 1934, Fig. 4), in spite of the fact that little schist is present in the western part.

(2) Ground originally mapped as Eilde Schist (Bailey, 1934, fig. 4), in and around the headwaters of a south-westerly draining stream on the southern side of Garbh Bheinn, has now been mapped as Eilde Quartzite. Here, there are good exposures of massive quartzite together with a conspicuous schist horizon. The schist is almost certainly a continuation of one of the several schist horizons, which are present within the Eilde Quartzite on the higher western slopes of Garbh Bheinn.

(3) The north-westerly trending Glencoe branch fault, shown on the 1 in. Geological Survey map as crossing Allt Gleann a' Chaolais just before its north-westerly termination, is now regarded as more probably remaining entirely on the south-westerly bank of the stream.

(4) A north-easterly trending fault, hereafter called the Torran nan Crann fault, has been recognized in the ground to the south-east of Torran nan Crann. A short distance downstream a second fault is actually exposed and in all probability it is a minor branch of the fault just mentioned. Both faults are believed to terminate against the Glencoe branch fault.

Lithological Types

(*a*) *Eilde Quartzite and Binnein Quartzite.*—The Eilde Quartzite, as developed around Loch Leven, is nearly always feldspathic, and in this respect it differs from the much purer Binnein Quartzite (Bailey, 1934, p. 490). Intercalated beds of schist are more commonly developed in the former, however, than have hitherto been recorded, and appear to reach their maximum development both on the higher western slopes of Garbh Bheinn, as well as on the south-eastern slopes adjacent to the junction with the Eilde Schist.

(*b*) *Eilde Schist.*—In the Memoir explaining Sheet 53 (p. 64), the Eilde Schist is described as a dark grey biotite schist nearly always containing quartzose intercalations up to about 2 feet in thickness. As far as the present area is concerned, however, it is possible to divide the Eilde Schist into two distinct facies depending upon the amount of quartzite present, and it is suggested that they should be called the schist facies and the banded facies respectively. In the vicinity of the fault intrusion the schist undergoes thermal metamorphism, but clearly this does not radically affect the distinguishing features of the two facies.

Schist facies.—This facies is a dark coloured biotite schist commonly containing insignificant quartzose ribs. Petrographically much of it is semipelitic in nature rather than pelitic.

Banded facies.—This facies consists of interbedded biotite schist and quartzite in very variable proportions. The individual beds vary widely

in thickness, so that the heterogeneous nature of this facies is not always apparent in a single small exposure.

Western Outcrop

(*a*) *Eilde Schist.*—Schist facies.—For about one mile southwards from Loch Leven by far the greater proportion of the Eilde Schist belongs to the banded facies, particularly good exposures being found on Torran nan Crann. The chief exception to this occurs on the southern side of the road, where the schist nearest to the junction with the Binnein Quartzite contains appreciable amounts of interbedded quartzite. The areal distribution of these latter beds, however, is too limited to be shown on the map of Text-fig. 1.

Only the middle beds of the schist facies persist unchanged as far as the Torran nan Crann fault, while the lower and upper beds change along the strike into the banded facies before reaching the fault. This transition, however, can only be fully demonstrated on the northern side of the stream draining the north-eastern side of Torran nan Crann, near to its source. Here, the transition is completed in about three hundred yards and is seen to be brought about by the increase of quartzite beds at the expense of the schist. Elsewhere the position of the transition zone can only be approximately inferred owing to the absence of good exposures.

Banded facies.—The Eilde Schist lying to the south-east of the Torran nan Crann fault belongs to the banded facies. The ground is not uniformly exposed, and the presence of a preponderating amount of schist in a few of the more conspicuous exposures may at first sight give the erroneous impression that one is dealing with the schist facies, but a detailed survey of the whole area leaves no doubt that a considerable amount of interbedded quartzite is present. In a few places the outcrops consist almost entirely of massive quartzite, as for instance in the upper part of the stream which is cut by the Torran nan Crann fault.

(*b*) *Top and bottom junction of the Eilde Schist.*—Throughout much of its length the junction between the Eilde Quartzite and the Eilde Schist appears to run parallel to the outcrops of the individual beds on either side of it. This is confirmed by some excellent exposures of the actual junction, which occur on the north-western side of Garbh Bheinn. Here there can be no doubt that the Eilde Quartzite shows an interbedded junction with the banded facies of the Eilde Schist.

In the vicinity of the fault intrusion, however, it is difficult to be absolutely certain whether such a simple relationship still persists, since the tracing of both the junction as well as the individual beds is complicated by the fact that the strike of the beds is occasionally subject to rapid local variations. As will be apparent from a later

paragraph, it is not inconceivable that this part of the junction is actually diachronic. Unfortunately the extreme southern end of this junction, just before it is cut off by the fault intrusion, lies in very poorly exposed ground, and consequently it provides no further information on the matter.

Except for the exposures in Allt Gleann a' Chaolais there are very few exposures of Binnein Quartzite east of the stream, and these are almost entirely found within the first quarter of a mile on the southern side of the road. In this latter locality there is considerable overfolding both of the Binnein Quartzite as well as the Eilde Schist, the pitch of the folds being gentle and generally directed either towards the north-west or the south-east. The folds are commonly recumbent in form and one well exposed example in the quartzite has an amplitude of almost 50 feet.

From the limited evidence available it would appear that the Binnein Quartzite follows the Eilde Schist conformably, but it should be pointed out that this may not necessarily be the case. In fact, there may actually be an undetected surface of erosion at the top of the Eilde Schist, similar to that which is believed to be present along the eastern outcrop. If this is correct it might possibly account for the apparent increase in thickness of the Eilde Schist towards the Torran nan Crann fault.

Eastern Outcrop

(*a*) *Eilde Schist.*—Lithology.—The whole of the eastern outcrop, including the outcrop within the fault intrusion, is probably best regarded as belonging to the banded facies. In general the Garbh Bheinn outcrop contains more schist and less quartzite than the banded facies on the western side of the hill, and it is only the small quartzite-rich fault intrusion outcrop which is strictly comparable. Nevertheless, there is an ill-defined quartzite-rich zone on the northern face of Garbh Bheinn adjacent to the Eilde Quartzite. Owing to the lack of exposures, however, it is not possible to say whether such a zone extends north of the road as well as down the southern face of the hill.

Structure.—The structure of the Eilde Schist is probably very simple and, as far as the present area is concerned, there is no support for J. F. N. Green's statement (1931, p. 529) that the rocks around Loch Leven are " lying in sheets of packed folds ". Internal folding is not entirely absent, however, as may be seen by reference to Text-fig. 1.

Only the anticlinal part of this double fold is exposed, and it is seen to have a moderate southerly pitch, while the position of the complementary syncline on the east has to be inferred. The folding definitely does not persist very far southwards and probably dies out in this direction. What happens northwards is even more problematical

owing to the absence of suitable schist exposures but, since no trace of such folding is found in the adjacent Binnein Quartzite, it probably dies out in this direction as diagrammatically shown in Text-fig. 1. Further reference to this folding will be made in a later paragraph.

(*b*) *Junction of Eilde Quartzite and Eilde Schist.*—On the northern face of Garbh Bheinn up to about 2,000 feet the junction between the Eilde Quartzite and the Eilde Schist runs parallel to the outcrops of the individual beds on either side of it. The discrepancy between the junction and the strike of the beds, seen in Text-fig. 1, is probably entirely due to the incidence of topography upon the obliquely inclined beds. Good exposures of this part of the junction are seen in some conspicuous crags at a height of about 2,000 feet, and as far as can be seen the Eilde Schist follows the Eilde Quartzite conformably.

Immediately above these crags and extending over the crest of the hill the junction is markedly diachronic, but unfortunately the greater

TABLE I

APPROXIMATE THICKNESSES OF BEDS AT THE LOCALITIES SHOWN IN TEXT-FIG. 1

EILDE QUARTZITE		EILDE SCHIST	
Locality	Thickness	Locality	Thickness
			ft.
U	Possibly about 100 feet	W	1,700
V	1,450 feet	X	1,100
Z	At least 450 feet	Y	800

part of it lies in poorly exposed ground. The strike of the beds on either side of it, however, appears to be the same. The only exposure of the junction in this vicinity is adjacent to the fault intrusion, where it can be seen for a dozen or more yards. Here it has been intruded by a thin porphyrite sill, which also cuts the fault intrusion. In spite of the extent of the exposure it shows no signs of being diachronic, while the contiguous beds are seen to have the same dip and strike. Lithologically there is an abrupt change between the massive quartzite on one side, which is very feldspathic in this area, and the schist on the other side.

The diachronic part of the junction concides with a marked variation in thickness of the contiguous beds. Traced from north to south, the Eilde Quartzite remains relatively thin at first, but near the summit of the hill and down the southern side a rapid increase in thickness takes place, ascribed to the incoming of additional beds at structurally lower levels. *Pari passu* with this change, the Eilde Schist shows a

correspondingly marked decrease in thickness when traced southwards from locality W in Text-fig. 1.

Some idea of the order of magnitude of these thickness variations may be obtained from Table I, which gives, very approximately, the thicknesses of the beds at the localities marked U to Z in Text-fig. 1. The figures for the Eilde Quartzite are not original thicknesses, but refer to the succession preserved in the fold, where there are no structurally higher beds to form an upper limit. The figures for the schist are calculated on the assumption that there is no more large scale internal folding than is shown in Text-fig. 1.

Before its destruction by the fault intrusion the junction between the Eilde Quartzite and the Eilde Schist south of Garbh Bheinn was probably also diachronic, since the thickness of the Eilde Schist is very much less in the fault intrusion outcrop than when last seen on Garbh Bheinn. Indeed, prior to the emplacement of the fault intrusion, the thickness of schist in the fault intrusion outcrop could not have been much more than 250 feet. No reliable figure can be given for the original thickness of quartzite on this eastern limb of the isoclinal fold, but calculations suggest that it was probably somewhat thicker than on the southern slopes of Garbh Bheinn.

The junction is exposed at intervals along the fault intrusion outcrop but, contrary to expectation, it is definitely not diachronic. As in the case of the exposure on the southern slopes of Garbh Bheinn the beds on either side appear to have the same dip and strike.

Nature of the junction—Any theory put forward to explain the nature of the junction between the Eilde Quartzite and the Eilde Schist, along the eastern outcrop, must not only take into account the variation in thickness of the quartzite belonging to this outcrop but also the variation throughout the whole area, since all these problems are probably related to one another. One possibility which satisfies these requirements, is to regard the junction of both outcrops as an unconformity, but clearly this conflicts with the evidence of the actual exposures. Of course, it may be argued that it is purely a coincidence that these exposures show parallelism of the beds, and that in each case there is an undetected stratigraphical break. But this is highly improbable.

A more promising theory, however, is to regard the diachronic part of the junction as traversing an area across which there is a rapid facies change from Eilde Quartzite to Eilde Schist. Not only does this account for all the thickness differences of the beds, but it also has the advantage of providing a simple explanation of the apparent conformability of the Eilde Schist, even in exposures closely associated with the diachronic junction.

Although the very poorly exposed nature of the ground makes it

impossible to observe such a facies change directly, there are a number of lithological observations which, whilst not proving its existence, at least increase its probability. For instance, both stratigraphcal units contain quartzite and schist, though in different proportions, while the beginning of a facies change may possibly be indicated by the marginal quartz-rich zone, already noted as being present in the Eilde Schist on the north-eastern side as well as possibly on the southern side of Garbh Bheinn. A preponderance of quartzite in the fault intrusion outcrop of the Eilde Schist, in comparison with the Garbh Bheinn outcrop, also points to a facies change. Further supporting evidence is supplied as follows by two beds of schist in the fault intrusion outcrop :—

(1) The thick bed of schist, which is regarded as marking the western junction of the Eilde Schist, splits up into several smaller beds with intervening quartzite, when traced towards the south. Had it not been for the southerly intervention of the fault intrusion it is probable that the schist beds would have thinned out altogether, thus completing the transition of the larger bed into quartzite indistinguishable from Eilde Quartzite.

(2) Within the Eilde Quartzite and about twenty yards from its junction with the Eilde Schist there is a bed of schist 2 feet thick. Traced southwards over continuously exposed ground, this bed shows a rapid transition into feldspathic quartzite. In all probability this bed once thickened towards the north, across the ground now occupied by the fault intrusion, and at some stage its prominence would have necessitated its inclusion within the Eilde Schist.

Although a few schist beds, including one about 15 feet thick, occur on the south-eastern slopes of Garbh Bheinn, lack of exposures of the Eilde Quartzite together with the intervention of the fault intrusion unfortunately prevent their being traced far enough for them to provide any information about the postulated facies change.

Tectonic significance of the facies change.—In offering an explanation for the fact that there is a smaller thickness of Eilde Quartzite in the north-eastern limb of the Garbh Bheinn downfold than in the south-western limb, Bailey (1934, p. 485) said that " it is probable that there has been some sliding ". It is now suggested that this difference in thickness can be largely or even wholly explained by the underground continuation of the above-mentioned facies change. Thus the " missing " quartzite of the north-eastern limb is possibly represented by part of the Eilde Schist as diagrammatically shown by the section in Text-fig. 1. A similar explanation would also apply to the contrast in thickness between the quartzite south-east of locality Z and that at locality V.

Finally, if these explanations are correct, it might be expected that the western junction between the Eilde Quartzite and the Eilde Schist

might also show some signs of being diachronic. The only part of this junction, however, which might conceivably behave in this way, as already tentatively suggested, is the somewhat problematical part in the vicinity of the fault intrusion.

(*c*) *Junction of Eilde Schist and Binnein Quartzite.*—The relationship of the junction between the Eilde Schist and the Binnein Quartzite to the contiguous beds varies from place to place. In the vicinity of the road and the loch, where the strike is rather variable, it maps out as a conformable junction and yet, for a considerable part of its length south of the road, it is markedly diachronic. This is somewhat diagrammatically shown in Text-fig. 1. Then, south of the noticeable bend in the junction, it is slightly diachronic for several hundred yards (not shown in Text-fig. 1), and finally, as far as can be seen from the limited exposures of the Binnein Quartzite, it once again appears to behave as if it were conformable. The entire junction is unexposed, except along the shore of the loch.

Before discussing the exact nature of the junction it is of interest to note that the diachronic part of it coincides with marked variations in thickness of the contiguous beds. The Eilde Schist, which is 1,700 feet thick at locality W as already mentioned, is considerably less along the line of section (Text-fig. 1), while in the vicinity of the road it is only about 600 feet thick. The Binnein Quartzite, on the other hand, shows a rapid north-westerly increase in thickness from a negligible figure, where the outcrop is narrowest, to very approximately 1,000 feet in the vicinity of the road. This increase is clearly due to the incoming of older beds towards the north-west.

Nature of the junction.—The diachronic nature of part of the junction suggests the possibility that, once again, we are dealing with a facies change problem. Such a possibility, however, is quite definitely ruled out by the lithological evidence, since the Binnein Quartzite, even in exposures relatively near to the junction, contains no interbedded schist.

The only theory which fits the facts reasonably well is to regard the top of the Eilde Schist as an erosional surface. In this case the behaviour of the northern part of the Binnein Quartzite can be ascribed to overlap, resulting from the beds having been deposited upon an irregularly eroded floor. The last part of the floor to have been covered by the quartzite must thus be represented by the southern part of the junction, which consequently must represent a disconformity. The increased thickness of quartzite in the extreme south of the area, as suggested by the increased width of outcrop, is believed to be due to the incoming of still younger beds not represented further north. The northern part of the junction near the road, where the postulated erosion is greatest, must also represent a disconformity.

Here, the basal beds of the massive Binnein Quartzite are locally somewhat feldspathic and contain a few thin beds of schist. It must be admitted, however, that although there are more or less continuous exposures along the shore of the loch from the Eilde Schist into the Binnein Quartzite, the exact position of the disconformity is still in doubt.

Finally, further mention must now be made to the folding in the Eilde Schist shown in Text-fig. 1, which clearly takes on a special interest in the event of the top of the schist being an erosional surface. Although the folding has been regarded as being closely connected with the main folding of the whole area, it should be pointed out that it is equally probable that it took place before the Binnein Quartzite was deposited. However, until further work establishes definite evidence of such early folding elsewhere, it is probably better to adopt the former interpretation.

Discussion and Conclusions

It might be expected that similar stratigraphical features to those just described should occur in other areas of the Lochaber Series, and yet the published work has little to say on the matter. Indeed, Bailey (1916, p. 81) remarks that " the facies of individual groups remains wonderfully constant ". Carruthers (1923, p. 26), however, briefly refers to the rapid northerly thinning of the Binnein Quartzite of Stob Ban (Sheet 54) as being due to " irregular sedimentation ", but he does not make it clear whether he is referring to a change of facies or to a lensing out of the deposit caused by lack of sedimentation. More recently J. G. C. Anderson has discovered an important case of a facies change in the Glen Roy district. Reference to this was made in the course of a lecture delivered to Section C of the British Association at the 1950 Birmingham Meeting, when Anderson drew attention to the fact that the whole of the Eilde Quartzite of this district shows a gradual north-easterly transition into Eilde Flags, so that eventually the Eilde Schist shows an interbedded junction with the Eilde Flags without intervening quartzite.

Furthermore, in the course of reconnaissance work north of Loch Leven, the writer has recently discovered yet another example of a facies change, this time in the Binnein Quartzite immediately north of Caolasnacon. When traced in a north-easterly direction it was found that the massive white quartzite develops intercalations of schist, which rapidly thicken at the expense of the quartzite, so that in the space of a few hundred yards the entire Binnein Quartzite is represented, not by quartzite, but by schist.

Not only is this example of interest because it closely parallels the

postulated relationship of the Eilde Quartzite and the Eilde Schist on the south side of the loch, but also because it leads to a very different interpretation of the ground than is indicated on the 1 in. Geological Survey map. According to Bailey (1934, p. 506) the north-easterly disappearance of the Binnein Quartzite is due to the intervention of a slide, which cuts out the quartzite so that the flanking Binnein Schist and Eilde Schist are brought together.

It will now be of interest to compare briefly the area described in this paper with the immediately adjacent area north of Loch Leven, where there is a continuation of the Garbh Bheinn downfold structure. According to Bailey (1934, p. 489) sliding is responsible for the attenuation of the Eilde Quartzite belonging to the eastern outcrop north of the loch, but it is now suggested that a very different explanation is possible. As on Garbh Bheinn, the " missing' ' quartzite may also be represented by part of the Eilde Schist, due to an underground facies change. The existence of such a facies change is not as fantastic as it might appear at first sight, since the unevenly serrated southern part of the junction between the quartzite and the schist along the western outcrop, as shown on the 1 in. Geological Survey map, may represent the surface manifestation of such a change.

Judging by the narrow width of the eastern outcrop of the Eilde Schist, it would appear that the schist is considerably attenuated in thickness for several miles north of Loch Leven. If this is true, then it is possibly due to a northerly continuation of the erosional surface at the base of the Binnein Quartzite, already postulated south of the loch.

Further reference to the 1 in. Geological Survey map shows that the junction between the Binnein Quartzite and the Binnein Schist of the eastern outcrop has been mapped as a slide. It is therefore of considerable interest that, although work on this junction is not yet complete, there are already indications that it is probably unnecessary to postulate any sliding. As far as can be ascertained the only reason for introducing sliding on the southern side of Loch Leven is to account for the thickness behaviour of the Binnein Quartzite but, as already explained, the thickness variation on the northern side of Garbh Bheinn is possibly due to sedimentary causes.

In conclusion, it thus appears that the stratigraphy of the Lochaber Series is more complicated than has hitherto been recorded, and that the existence of facies changes and erosional surfaces makes it unnecessary to postulate minor sliding to account for the rapid thickness variations of the beds. A tentative comparison of this small area with the area north of Loch Leven appears to substantiate these conclusions, and it is therefore suggested that a re-examination of the whole series would be well worth while.

ACKNOWLEDGMENTS

The writer wishes to acknowledge his grateful thanks to Sir Edward Bailey for introducing him to the structural geology of the area, and for many helpful discussions during the last ten years. He is also indebted to Professor F. W. Shotton for his guidance on a number of points during the writing of the paper.

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DEPT. OF GEOLOGY,
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BIRMINGHAM, 15.

Carbonatite Structure : Tororo Hills, Eastern Uganda

By C. E. Williams

Abstract

The carbonatite plugs of Eastern Uganda represent the final phase of Elgon type volcanoes. The types of structure observed in the carbonatites are discussed and a general description of the structural components of the Tororo Hills is given. It is suggested that the ring features originated through the congealing in the conduit of the central portion of the carbonate mass.

RESULTS of work done in Norway, South Africa, and Uganda, suggest that carbonatite complexes are likely sources of various economic and strategic minerals. For this reason it is felt that some attempt should be made to record the structural features observed.

The Tororo Hills are situated on the extreme eastern boundary of Uganda, near the main Mombasa–Kampala railway line. The most striking feature is the plug-like Tororo Rock which stands 800 feet above the surrounding country (Text-fig. 1). The Rock and the lesser hills of the group are made up of sheeted alvikitic sövites.

The Tororo occurrence is related, genetically, to the larger mass of carbonate rock in the Sukulu Hills, 2½ miles to the south-west. It is hoped that the Sukulu group will form the basis of a more comprehensive study later.

It is clear from the work of K. A. Davies (1951) on Elgon and B. C. King (1948) on Napak that the carbonatite plugs of Eastern Uganda represent the final phase in the history of volcanoes of the Elgon type.

The remnants of the cones of Napak and Mt. Elgon are made up of agglomerates, tuffs, mud flows, and lavas of phonolitic and nephelinitic composition. These rocks indicate the extremely explosive nature of the occurrence and the vast amounts of gas and water liberated.

At Napak the " punch " of the volcano tilted the basement granites outwards for a radius of over 4 miles. The original conduit is plugged by carbonate rocks which form a very small percentage of the whole mountain.

At Tororo and Sukulu only the carbonate masses remain to indicate the former presence of a large volcano, but diamond-drilling has proved the presence, if not the extent, of fenite rings around the plugs.

The country rock is, in all observed cases, granitic and no limestones, of an origin other than volcanic, have been recorded in the region.

Several types of structural features have been observed in the carbonate core :—

1. The massive features which are visible at a distance. These

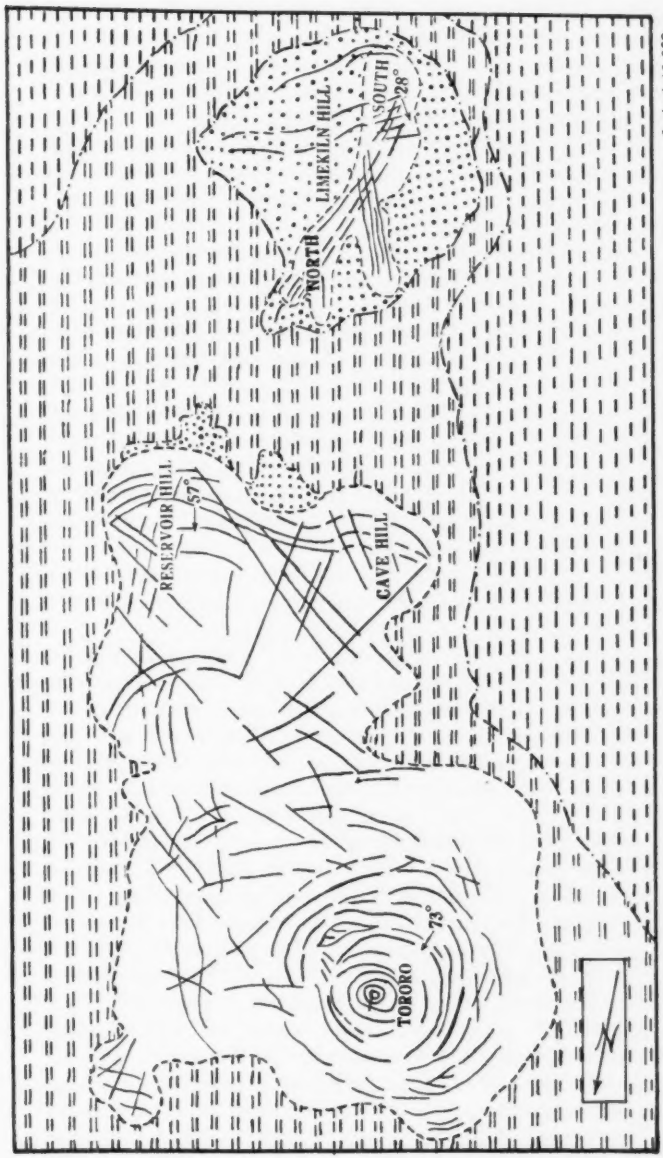

FIGURE 1 Simplified trend diagram of Tororo Hills

take the form of " cone-sheets ", dykes, and ring-dykes, and although they are composite bodies tend to stand out as marked units, characterized by sheer faces covered by deposits of iron-stained secondary limestone. In many instances they have the appearance of sedimentary strata dipping inwards towards the centre.

The main volcanic neck at Tororo is encircled by one of these bodies which is in turn circumscribed by a further ring-structure of greater diameter. It is suggested that the term " collar structure " might usefully be applied.

2. Detailed examination of the massive features reveals the presence of lamellar flow structures generally agreeing with the large-scale components. Some divergence is found where factors such as faulting and shearing have induced a secondary lineation.

Turbulent flow is recorded around wall rock inclusions and is general in the fenite areas.

3. The most obvious of the detailed structures are the innumerable fine-grained dykes cutting the carbonatite. These vary in width from a fraction of an inch to about one foot. They have been intruded in several generations, probably under conditions of tension during periods of structural readjustment, the earlier sets being displaced by the later ones.

On outcrop these dyke-swarms are yellow-brown but at depth are grey-white and contain over ninety per cent calcium carbonate. The dyke walls show clear evidence of dilation.

4. In addition to the small-scale faulting associated with the dyke-sets, there are also large radial faults which are younger than some of the outermost sheets. This type of faulting is more marked on Sukulu where subsidiary, later centres have developed along some of the fault lines. These faults are characterized by considerable horizontal displacement and intense brecciation, giving rise to sövite breccias and fenite-breccia dykes, the latter indicating explosive intrusion along the structural weakness.

5. It has been noted that while the major faulting radiates from the intrusion centre, very few of the dykes are true radials. They are almost invariably tangential. Close examination indicates that these dykes originate as part of the ring structure but tend to straighten out, particularly along the grain of the country rock. This tangential aberration of ring structure, which has an important bearing on economic considerations, has been observed only in the high angle sheets fairly close to the centre of the plug. In general, they tend to retain something of the inward dip until the sphere of influence of the centre is escaped when the primary lineation becomes vertical. It is, therefore, believed that these features run out laterally and do not necessarily arrive in their

present position from directly below. This is certainly the case on Lime-kiln Hill where diamond-drilling has shown that the limestone body is floored by syenitic material.

6. The jointing in the sövite is generally of the exfoliation type, the wide joints being filled with coarsely crystalline calcite. Since this jointing is parallel to the surface, it is only seen in recently opened excavations. A further joint system is orientated parallel to the lineation.

7. The weathering of the carbonate rock tends to accentuate the structural features. The massive components stand out because of their greater resistance to weathering than the adjacent sheets. The lineation is brought into sharp relief through fluting and the dyke-swarms made conspicuous by their discoloration on weathering. Dolomitic segregations are obvious because of their typical weathering. A curious feature of the carbonate rock is its ability to resist weathering to the extent that it stands 300 to 800 feet above its fenite pediment. The answer probably lies in the fact that while the sövites are reasonably homogeneous, the fenites are complex mixtures of the various rock types with minerals more liable to hydration and oxidation than those of the carbonatite.

It may be mentioned here that extremely high yields are obtained from water bore-holes sited in the mixed rock zone surrounding the carbonatite.

Description of Tororo Hill Structure

The central part of the Tororo plug is a small structureless mass, about 10 feet across, of medium-grained, red-stained sövite. This is cut through the centre by a small calcitic dyke. Immediately surround- the inner body is a series of rings of fine-grained, limonite-stained sövite displaying vertical lineation. This occupies most of the top of the hill.

With the step down on to the first of the collars, the sövite becomes coarse-grained and the sheets are inclined very slightly towards the centre. As the collars are traversed outwards from the centre there is a gradual decrease in the angle of dip until, at the foot of the Rock, a dip of 73° is recorded.

From the foot of Tororo Rock the complex develops an eccentricity towards the south, in which direction are three very much smaller hills.

The trend of tangential aberration of ring structure at both Sukulu and Tororo follows an anticyclonic pattern with departure from the ring structure generally taking place in an anticlockwise manner. The two hills, known as Reservoir Hill and Cave Hill respectively, show the result to be expected when clockwise and anticlockwise

features converge. The structure on Reservoir Hill is that of a tangential dyke arising from the eastern foot of Tororo Rock while the Cave Hill structure is that of a tangential dyke originating on the western foot. The outbreak of the two components was clearly simultaneous, for the lineation of each cuts the other indiscriminately. A series of collar sheets dipping inwards at 57° form the basis of the outward termination of both the Reservoir Hill and Cave Hill. These are older than the tangential dykes which pass through them.

The Limekiln Hill may be divided into two separate parts, the North Kiln Hill, which is an extension of the tangential dyke from Reservoir Hill, and the South Kiln Hill which, although originating from the same source as the former, is separated from it by the strongly cross-cutting components from the Cave Hill dyke.

The North Kiln Hill is composed of alvikitic sövite contaminated by fragments of syenitic material, while the South Kiln Hill, where the carbonatite is of the same composition, has fewer xenoliths.

The Limekiln Hill carbonatite is contained within a narrow trough, the walls of which are composed of shattered fenites in process of digestion by the sövite. As already mentioned, the intrusion is floored by syenite. This floor is inclined towards the point of aberration of the western tangential dyke. The primary lineation of the dyke is vertical and parallel to the sides of the trough except where xenoliths have caused turbulent flow. The dyke is modified by the later collar sheets which have been intruded at a low angle parallel to the syenite floor. Once again the axis of intrusion is directed towards a point west of Tororo Rock external to the centre. From the intense brecciation of the walls and distal end of the trough, it is clear that this body became emplaced under explosive conditions. Parallel to the flow lamellae are a series of narrow pyroxene-breccia dykes which are slightly older than the collars.

It is proposed to use the South Kiln carbonatite in the manufacture of cement.

At both Tororo and Sukulu it is clear that the individual so-called " cone-sheets " increase in angle as they approach the centre. This has also been noted by Strauss and Truter (1950, p. 108) in Sekukuniland. The occurrence, therefore, takes the form, in the ideal case, of a funnel with a parabolic profile. In the field the term " collar structure " is used to differentiate these bodies from true cone-sheets. The use of the latter term is liable to induce fallacious arguments as to the depth of the intrusion loci. In view of the tremendous " punch " known to be associated with these volcanoes, it is likely that a cone of percussion would be produced. The possibility of the existence, therefore, of true cone-sheets is not excluded but it is thought that these would be intruded earlier in the history of the volcano,

during periods when the vent was choked with extrusive material and that the rock types in the true cone-sheets would be representative of the flow materials of the earlier phases.

It is believed, on evidence found at Tororo, Sukulu, and other alkaline centres in Eastern Uganda, that the central portion of the carbonatite plug congealed in the volcanic conduit as a result of rapid escape of CO_2 leading to a drop in both temperature and pressure. It is possible that this occurred more than once in the history of the volcano and may possibly be regarded as the end of a single phase. Some of the Elgon agglomerates are known to contain fragments of carbonatite.

We know, however, that plugs at present occupying the conduits represent a phase in the history of the volcano when insufficient explosive force remained to enable the vent to clear itself of a large obstruction. Any remaining carbonate below the plug would then be intruded as ring features along the only possible zone of weakness, that existing between the congealing mass and the wall rock. Should any of the earlier cone-sheets actually pass the obstructing mass, they would tend to flow inwards over it. This might explain the outward dipping primary lineation recorded by Strauss and Truter (1950) at Spitskop.

It is thought that the major processes of fenitization took place after the obstruction of the conduit and that the fenite aureole was enlarged by the penetration of sövite into the country rock. Any mobilization of fenite material would occur at this stage.

The hypothesis, generally accepted, is that the innermost mass is the youngest. This presupposes that the vent was occupied by the congealed lavas, for without some obstruction or constriction, there would be no impetus to ring intrusion. It would therefore be expected that xenoliths in the carbonatite would represent the lava types. This is, however, not the case, and all inclusions so far observed are fenite fragments normally found in the area surrounding the carbonate.

Several other features contribute towards the hypothesis that sövite ring structure is the result of carbonatite obstruction of the vent.

Firstly, the central sövites are generally of finer grain than the outer sövites which would indicate rapid cooling of the central mass and a more leisurely emplacement of the outer rings.

The tangential aberration of ring structure follows an anticyclonic pattern, indicating a rotatory outward flow from the centre of pressure. The tangential dykes originate at a point away from the centre suggesting that the vent had already been plugged. A good example of a tangential dyke originating from the periphery of a volcanic plug may be seen in M. P. Billings' *Structural Geology*, Plate XVII, facing

p. 243. While it is not suggested that the feature shown here is an aberrant ring structure, it is thought likely that as the conduit became plugged with volcanic materials, the dyke material was forced to take a path between the plug and the wall rock.

As mentioned earlier, the tangential features of the Tororo and Sukulu Hills have a bearing on the economic development now in process in the area. It has been found, as a result of many thousands of feet of diamond-drilling, that the only bodies sufficiently low in magnesia for the manufacture of cement are the tangential dykes. The carbonate rock carrying the highest percentage of pyrochlore so far discovered is in the form of a tangential dyke. It is not true, however, to say that all tangential dykes carry high niobium percentages.

In conclusion, the author would like to point out that this paper has been written under field conditions without easy access to standard publications and before the completion of structural mapping in the more extensive Sukulu Hills. It is, therefore, more than likely that some modification of the views expressed here will be brought forward at a later date.

He would also like to express his sincere appreciation of the sympathetic guidance of Dr. K. A. Davies, O.B.E., Director of Geological Survey, Uganda, who suggested and permitted the publication of this paper.

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GEOLOGICAL SURVEY OF UGANDA,
P.O. BOX 9,
ENTEBBE,
UGANDA.

Volcanic Ash from the Hekla Eruption of 1947

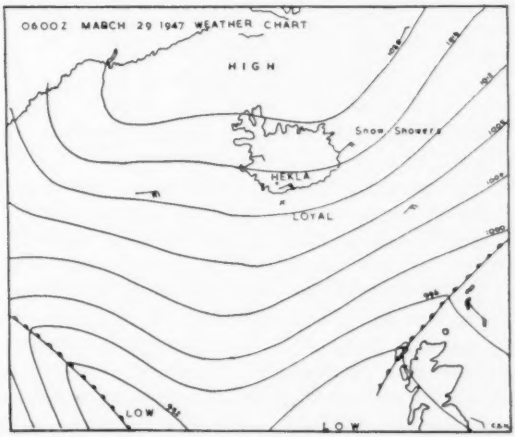
By Frederick A. Henson

Abstract

Description of ash which fell on a trawler, fishing off Iceland, during the first hours of the 1947 Hekla eruption. Discussion of the probable track of the ash from the volcano to the ship and the factors affecting sorting involving upper wind velocities, terminal velocities of the various sizes of pumice and mineral fragments which constitute the ash, and weather conditions.

Introduction

JUST over five years ago, at 6.50 a.m. on the morning of 29th March, 1947, the Icelandic volcano Hekla erupted after one hundred and two years of quiescence. New lavas were extruded from fissure eruptions, and vast quantities of ash and pumice were ejected in the first day of the eruption. Most of the ash fell over the southern coastal



Text-fig. 1.—Map showing weather conditions and the positions of Hekla and S.T. *Loyal.*

area of Iceland and in the sea. Dr. Tyrrell (*Nature*, May, 1947, and January, 1948) gave a description of the eruptions and new lavas.

The S.T. *Loyal* (Skipper S. Thorsteinsson) was fishing off the southern coast of Iceland at the time in a position Lat. 63° 22′ 15″ N. and Long. 19° 20′ 00″ W. This position is thirty-nine nautical miles from Hekla, the bearing of which is 346°. The crew of this Grimsby trawler saw the first of the eruptions and were in the path of falling

ash for four hours. A good representative sample of the ash which fell during this time was collected.

Description of the Ash

Macroscopically the ash which fell on the S.T. *Loyal* consists mainly of grey and occasional blue-grey coloured fragments of fine pumice. These fragments are angular, sometimes contorted and exhibiting elongation with sub-parallel air cavities. The whole of the available material was mechanically sieved for a short period to find its mechanical composition. The tabulated results given below are expressed graphically as a cumulative log. curve in Fig. 2. Each fraction obtained by sieving was further used to give the specific gravity of that fraction and for microscopic examination. The values of the specific gravities are also plotted on Fig. 2.

Results of Mechanical Analysis

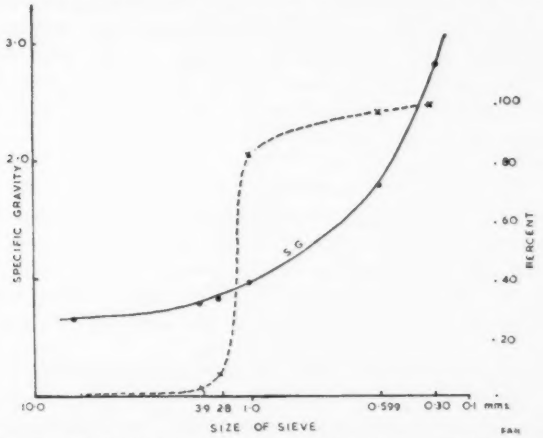
Size of Sieve mm.	*Weight %*	*Specific Gravity*
9·51	0·55	0·74
3·962	1·58	0·80
2·794	5·50	0·83
1·003	76·78	1·04
0·599	13·62	1·81
0·302	1·97	2·85

The cumulative log. curve in Fig. 2 shows that the ash is well graded, with the bulk (over 76 per cent) between 1·00 and 2·80 mm. in diameter. The coarser grained constituents are only approximately half the total weight per cent of the finer grained material forming the ash ; but, owing to the difference in their respective specific gravities, they are approximately volumetrically equal. The specific gravity values of the fractions, particularly those of the coarser ones, must only be regarded as approximate for two reasons. Firstly, in such heterogeneous material the size and degree of continuity of vesicular structures in the fragments are extremely variable. Secondly, since the fragments were not wax coated on account of their size, some of the outer cavities would be partially filled with water, thus giving a slightly higher value to the specific gravity. Furthermore, the values plotted in Fig. 2 represent the average for all constituents present. The values obtained, however, clearly lie on the curve shown above.

For microscopic examination all the fractions, with the exception of the finest material which was mounted directly in Canada Balsam, were first impregnated with shellac and then mixed with shellac in shallow cylindrical moulds for sectioning. In the sections obtained, where shellac is the mounting material, two quite distinct types of slaggy pumice occur in all fractions. The most abundant variety is compact in appearance and contains abundant iron ores, principally magnetite.

The abundant vesicles are almost spherical or slightly elongated. Less common is pumice in which the vesicles are much longer, considerably distorted, and exhibit a sub-parallel orientation. In this type, the degree of impregnation by the mounting material is more complete than in the commoner type. The amount of magnetite is variable ; it is often more abundant in the compact variety of pumice. Apart from magnetite and plagioclase-felspar microliths (often perfectly zoned) the pumiceous material is predominantly of light brown glass, with few recognizable mineral fragments, other than olivine.

The finest fraction, i.e. that retained by the 0·302 mm. sieve, is almost wholly crystalline consisting of abundant felspars, some



TEXT-FIG. 2.—Mechanical analysis of volcanic ash from Hekla eruption of 1947.

olivines, pyroxene, and magnetite. All these constituents occur as perfect euhedral crystals and in varying states of fragmentation ; they are unaltered but contain numerous inclusions.

The felspar present is labradorite (α 1·5587, β 1·5623, γ 1·5665, biaxial negative, An. 56 per cent) occurring as perfect euhedral crystals or crystal fragments showing fresh fracture surfaces. Under ordinary light these felspars exhibit perfect albite lamellar twinning, zoned structures, good cleavages and numerous inclusions. An unusual feature, exhibited on the surface of many of these felspars, is the development of light brown almost circular to polygonal markings. The surface markings may be caused by solvent action before ejection from the volcano, impressed vesicular markings, or may be due to sudden cooling. In addition to opaque inclusions of magnetite other inclusions, occurring both as perfect euhedral crystals and as hair-like

forms are common. Apatite is common as elongated prismatic crystals exhibiting poor basal cleavage. Within the small apatites minute rod-like material, similar to the enclosing apatite, may well represent earliest apatite. Rarely euhedral and sub-hedral olivine occurs within labradorite.

Olivine (α 1·698, β 1·718, and γ 1·739; chrysolite, 32 per cent $Fe SiO_3$) is abundant, being less common that labradorite but more frequent than hypersthene. This mineral occurs both as perfect and fragmented crystals. It is not surprising that olivine is extremely fresh and unaltered. Very occasionally the development of circular-polygonal markings on the surface occur. On many fragments a perfect conchoidal fracture, presumably developed during explosive fragmentation is seen. In colour the olivines are variable from light yellow to a brown iron stained colour. The olivine present is very slightly pleochroic. Inclusions are less common than in the felspars, but apatite in perfect prismatic crystals and hair-like rods, occurs as well as gaseous inclusions. The magnetite is mainly euhedral.

Hypersthene (α 1·6890, β 1·7023, γ 1·7156; 30 per cent $Fe SiO_3$) occurs as euhedral and sub-hedral bottle green crystals which are distinctly pleochroic and give straight extinction. The cleavage parallel to 110 is well developed in only a few crystals. Unlike labradorite and chrysolite the hypersthene contains no inclusions of apatite and the surfaces do not show polygonal markings. This mineral, however, does contain numerous gaseous cavities and magnetite inclusions.

Factors Affecting Deposition and Sorting

Surface winds recorded at the time of the eruption, and in the following six hours, were mainly light north to north-easterly. The gradient wind from the weather chart is 030° 15 to 20 knots. It is clear, therefore, that the ash was brought to the ship's position (south-south-east of Hekla) by upper winds from the north, or even the west-north-west. The photograph, accompanying Dr. Tyrrell's paper in *Nature*, shows an immense cloud of steam and ash over Hekla, the upper part of which is moving from north-west to south-east. The height of this north-westerly upper wind is judged to be in the region of 25,000 to 30,000 feet (mainly on account of the cumuliform nature of the cloud and the development of false cirrus). The upper winds from the 500 millibar daily weather chart appear to have been of the order of 30 to 35 miles per hour at the time of the eruption, and to have increased and backed slightly by midday. On the basis of these wind velocities the track taken by the ash would appear to have been a semi-circle of 60 to 70 miles; the average ground speed of the ash being 35 to 40 miles per hour.

The initial column of steam and ash from Hekla rose to over 20 km.

in height, according to eye witness accounts, whilst a wall of black smoke and ash 10 km. high extended over southern Iceland and out to sea. In order to reach the ship's position the ash from the volcano (which is assumed to have started falling earthwards in the vicinity of Hekla) must have taken from 90 to 120 minutes to reach sea-level. This means that if the particles of pumice soon attained their terminal velocities, then the average terminal velocity would be of the order of 5 to 6 feet per second. Such a figure is considered to be far too low and atmospheric turbulence and instability must have partially counterbalanced gravitational force.

To prove the validity of this argument the terminal velocities of the various sizes of pumice and fine crystals were found experimentally by balancing particles against a column of air moving upwards through a long glass tube. The results give an approximate range for the terminal velocities which cannot be obtained mathematically.

Size of particle. mm.	*T.V. Metres per minute.*	*T.V. feet per second.*
*9·51	350 +	Over 20
3·962	300–330	16·4–19·0
2·794	250–300	13·7–16·4
1·003	200–250	10·9–13·7
0·599	150–200	8·2–10·9
0·302	140–170	8·0– 9·3

* This value is very approximate.

Assuming all the particles to have started at the same height of 30,000 feet, then all particles of the sizes given above would reach sea-level in times ranging from twenty-five to sixty minutes. These times are much less than the time required for the particles to have been carried from Hekla to the trawler. Rising air currents, associated with showery conditions during the morning, delayed the particles in their descent. Since some particles have terminal velocities double the lowest values obtained and probably started to fall at lower altitudes than the smaller particles, it is suggested that increasing instability during the morning distributed the coarser material further afield. (Thunderstorms occurred on the south coasts at midday.) The distribution of ash, in the first few hours following this eruption, is concluded to have been mainly controlled by upper wind velocities and atmospheric instability.

Immediately after the Hekla eruptions the Danish people were asked to keep a look out for volcanic dust from Iceland. Very fine red dust, which fell at the end of March in Denmark, was described by Arne-Noe-Nygaard [1] and the author is indebted to Dr. G. W. Tyrrel for a copy of his description.

[1] Et. Stovfold over Danmark i Slutninger at Mars 1947.

This very fine red dust also fell in Frankfurt but it contained no volcanic material. It is considered to have been transported by upper winds from the Sahara.

The author wishes to express his thanks to Skipper S. Thorsteinsson, to Mr. Ives who brought the material into the Department, and to the Director of the Naval Weather Service for kindly supplying surface weather charts of the Iceland area.

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REVIEWS

INTERNAL CONSTITUTION OF THE EARTH. Second revised edition of Volume VII of the Physics of the Earth. Edited by R. GUTENBERG. Dover Publications, Inc., 1951. pp. 439, 43 text-figs. Price $6 (44*s.*).

This new edition of a familiar book first published in 1939 is written by eleven outstanding authorities on geophysics and includes all aspects, apart from terrestrial magnetism and electricity, of the deeper structure of the earth. For a work which summarizes such a wide field and which, with its extensive bibliography, will be a valuable reference book for several years, it is a pleasure to find that, in general, it is both coherent and readable, rather than disjointed with factual information and dry.

As with the first edition, and as may be expected with a book containing contributions from numerous authors, there are parts in which the general balance is not maintained. This occurs in those chapters where the authors attempt to provide too much detail rather than the references to, and general significance of, this detail. For example there are two chapters which between them constitute more than a quarter of the book and which contain about sixty pages of tabulated figures. With these two chapters it is disappointing to find that they carry forward a weakness of the first edition into the second by being among the six chapters out of sixteen which have not been rewritten.

The general quality of the book is, however, so high that its weakness is far outbalanced by its strength.

As with the first edition, the great burden of the work has fallen upon Professor Gutenberg. Not only has he edited the book but has written, in full or in part, nine chapters. He and the other writers must be congratulated for providing a work which will be of great value both to the specialists and to scientists in other fields.

M. N. H.

GEOLOGY OF THE APPLEBY DISTRICT (3rd edition, revised). By H. C. VERSEY. Whitehead, Appleby, 1951. pp. 40, 6 figs., 4 plates. Price 3*s.* 6*d.*

When the late Professor Gilligan wrote his handbook to the geology of the Appleby district he performed a valuable service to field workers by putting into their hands a concise guide to an area which presents simple geology on a grand scale side by side with problems of detailed complexity. His booklet became the inevitable companion of the hosts of students who were brought for training to this delightful piece of unspoiled country. In 1941 Dr. Versey undertook the revision of his late colleague's handbook, incorporating much that was new and adding itineraries of selected excursions. For a second edition in 1947 he had the expert help of Mr. J. Selwyn Turner in dealing with the Carboniferous succession, and this assistance has again been proffered in

the third edition under review. The result is a guide of the utmost value and one which will assuredly continue to be a " best-seller ".

Dr. Versey's knowledge of the district and his researches, particularly upon the New Red rocks, enable him to write with authority. Between his views on such subjects as the relative roles played during Permo-Trias times by the North Pennines, Lake District, and Eden Valley, and the opinions of some of the authors whose work he quotes, there have been and probably still are some fundamental differences. It is a tribute to Dr. Versey to say that he writes without over-emphasis of his own views.

The excellent written account is accompanied by good text-figures (though fig. 6 might have been acknowledged as reprinted from the *Quarterly Journal of the Geological Society*) and if any major criticism is offered, it concerns the half-tone illustrations. Though these give a good idea of the scenery the amount of geology which can be discerned is less than might be hoped for and, indeed, less than they purport to show. To some extent this is no doubt due to the screens being too fine for the quality of paper on which the photographs are printed, with the result that detail is often blurred ; but general views, even of rugged scenery, frequently show surprisingly little geology to anyone who does not know beforehand what he should see.

F. W. S.

THE TECTONICS OF MIDDLE NORTH AMERICA. Middle North America East of the Cordilleran System. By PHILIP B. KING. pp. xix + 203, with 52 diagrams. Princeton, New Jersey : Princeton University Press, 1951. 3.75 dollars. (London : Geoffrey Cumberledge. 24*s*.)

Within this slender volume the author has succeeded in presenting an admirable summary of the tectonics of the United States east of the Rocky Mountains. In view of the vast bulk of the literature on this subject, it is no small achievement to compress a readable and yet adequate account into the limits of 200 pages. The treatment is primarily descriptive, taking the structure of the country region by region, and although sufficient theory is introduced to unify the local details, it is never allowed to run ahead of the observed facts. Stratigraphical evidence is skilfully used in working out the details of tectonic development, and the reader incidentally obtains some very enlightening glimpses of the relationship between sedimentation and contemporary earth movement. Much trouble has clearly been taken with the illustrations, all of which have been redrawn to conform with a uniform scheme of presentation.

The author writes in simple and concise language, refreshingly free from redundant words, and, without any sacrifice of precision, he avoids the elaborate jargon that is all too apt to creep into modern writings on tectonics. " These terms," he says, " appear to have slight use in the present work, for distinctions can be expressed better by descriptive terms." And he amply justifies his claim.

The text is photographed from typescript, a style which never makes for easy reading, and some of the diagrams sadly over-reduced ; in these respects, the book unfortunately falls short of the standard of printing which we have come to expect from the Princeton University Press. Admittedly, this is only a preliminary edition : pressure of other duties has prevented the author from completing the chapters on the regions west of the Rocky Mountain front, and from preparing all the illustrations he intended ; but the book as it stands deserves better printing than this. We hope that when the work reaches completion its format will be more worthy of its contents.

In the meantime, the present edition can be warmly recommended to all who are interested in the geology of the United States, and in structural geology generally, for it is an important work, and provides a valuable guide both to the subject itself, and to its voluminous literature.

M. B.

SEDIMENTARY PETROGRAPHY. By HENRY B. MILNER. Third edition reprinted with a new preface and corrigenda. pp. xix + 666, with 100 figs. and 52 plates. London: Thomas Murby and Co., 1952. Price £2 10*s*.

We are pleased to know that copies of this useful work of reference are once more available, the publishers having reprinted the third edition with a new preface and a list of corrigenda. A fourth edition is in course of preparation, but is unlikely to be published for some little time to come, and a full review must await its appearance. In the meantime, we may remark that those parts of the book which a practical petrographer most frequently consults suffer least from want of revision, and *Sedimentary Petrography* retains its pre-eminence as a standard text on the minerals of sedimentary rocks.

M. B.

COAST EROSION AND PROTECTION: STUDIES IN CAUSES AND REMEDIES. By R. R. MINIKIN. 240 pp., with 226 figs. Chapman and Hall, London (for "Dock and Harbour Authority"), 1952. Price 30*s*.

The subject of this book lies on the borderline between physical geology and engineering, and has been of increasing importance since the early part of the century, when the publication of the Report of the Royal Commission on Coast Erosion focused attention on the complex issues involved. A fresh discussion of some of these problems is therefore welcome. It is evident from a perusal of Mr. Minikin's volume that he has made a long and detailed study of beach and foreshore phenomena. Yet the resulting impression left by the book is one of disappointment, due largely, perhaps, to the bewildering arrangement of the subject matter and the often discursive style of writing employed. Had the important general discussions been collected together and separated from the mass of descriptive detail, a more effective presentation would have resulted. Perhaps a partial attempt at this is found in Chapter 12 (headed "General Considerations"); but even here, after emphasizing the importance of a study of the interaction of waves and currents for a given foreshore, and commenting on the true function of groynes and the fundamental requirements of coast protection works, the short chapter closes with an anti-climax in the form of a section on the use of bitumen in maritime works. The feeling remains that by a little editing these unfortunate impressions could have been avoided.

Of the thirteen chapters which make up this volume of 240 pages, the first deals briefly with some aspects of geology, particularly with marine erosion, but it deserves more careful treatment and in places more accurate phrasing if it is to serve usefully to introduce geological ideas to non-geological readers. The tabulation of the mineral composition of common rocks (p. 3) is open to criticism. Chapters 2, 3, and 4 deal mainly with the erosion of cliffs and foreshores formed in different rocks, including glacial drift, clays, sands, and Chalk. It is unfortunate that short, overall headings to chapters are commonly lacking. Chapter 4, for example, is entitled "The restless change. Natural accretive phenomena. A coastline of variable erosive conditions. New Romney and Dymchurch sea walls. Complexity of beach phenomena"—a series of apparently disconnected topics which need a single connecting heading if they are to be grouped together. The fifth chapter: "The influence of stone jetties. Beach stability," begins with a discussion of eddy currents, and passes on to a description of groynes at Hythe and sea wall damage at Sandgate. Tidal and shore currents form the main subject of chapter 6. The three following chapters deal largely with the stability of beaches, including a section on littoral drift; on p. 118 ten factors in beach stability are listed, and these are used as main headings in this and subsequent chapters (though this is not stated in the text). Coast protection and types of protective works form the subjects of chapters 10 and 11. The succeeding chapter (General Considerations) has been mentioned above, and the last is an account of Hydro-graphical Surveys.

The book is illustrated by over 200 photographs and diagrams. Many of the photographs are of beaches, and some of these (for example, figs. 91, 92, 93 109) are poor, perhaps having lost in reproduction. No references to literature are given (except two on p. 223) although many opinions are quoted throughout the work.

F. G. H. B.

TABLES FOR MICROSCOPIC IDENTIFICATION OF ORE MINERALS. By W. UYTENBOGAARDT. Princeton University Press, 1952. (London: Geoffrey Cumberlege.) pp. 242. Price $5.00 (32*s.* 6*d.*).

Although the microscopical examination of transparent and of opaque crystalline matter began about a century ago, the study of the latter has lagged far behind that of the former. The difficulties, both theoretical and practical, of the study of minerals by reflected light are considerable, and the tables for systematic identification of ore minerals by this method are still in a very elementary state. When quantitative measurements of reflectivity are given (this cannot be done for one-fifth of the minerals listed in the present work), the method used (photometer ocular or photoelectric cell) has to be stated because the results often differ. The widespread existence of strong dispersion of reflectivity among ore minerals makes it very important to do these measurements in various monochromatic lights. It is regrettable that the author has seen fit to include, along with the figures for red, orange, and green light (taken mainly from the work of Schneiderhöhn and Ramdohr) some recent measurements made in white light only. Usable optical data in reflected light are so scanty in any case that it is a retrograde step to return to measurements in white light only.

In the present state of the data, the arrangement of minerals in comprehensive tables of this kind presents difficulties. The principal table here is given in order of "polishing hardness" mainly, although minerals which are closely related or which commonly occur together are taken together. The better the polishing, however, the more such differences of polishing hardness disappear, and the author suggests looking at sections before they get the final polish which is needed if quantitative measurements of reflectivity are to be made. He gives, also, a list in approximate order of increasing reflectivity.

The notes and references are very full and useful. The reproduction from Varitype is clear. It is only in use that the author's methods of presentation of the available data can be tested, but this book does mark a step forward towards the production of tables for use with reflected light which shall be as useful as those which we have for transmitted light.

N. F. M. H.

CORRESPONDENCE

THE GRANITIZATION PROCESS AND ITS LIMITATIONS

SIR,—In G. A. Joplin's recent paper, "The granitization process and its limitations" (*Geol. Mag.*, lxxxix, 25–38), the author has queried the possibility of the formation of basic fronts on a very large scale, except under "exceptional and rare" circumstances. It has been my good fortune to discover a basic front of considerable dimensions in south-west Tanganyika Territory, in the mountainous region of Ukinga, lying at the north-eastern end of Lake Nyasa, where, during what I have called the Ukinga Metamorphism, there occurred regional compression directed from the south-west, relieved mainly by a series of large thrusts. This thrusting forced up wedges of Basement System gneisses into younger sediments. The Basement System is dominantly amphibolitic, but some granitic gneisses were also involved in the dislocation. In many places the amphibolites were caused to over-ride the

folded, sheared, and mylonized sediments. Along one of the major dislocations amphibolites were thrust over sediments along a front of more than 50 miles.

Associated with the thrusting was an influx of potash, silica, and minor constituents including boron, beryllium, lithium, fluorine, chlorine, phosphorus, and carbon dioxide. This metasomatism gave rise to profound changes in the amphibolites, transforming them to acidic potash-permeation augen gneisses. The material displaced from the amphibolites, largely iron and magnesia, was driven out into the sediments, producing, from mudstones and siltstones, a group of chlorite permeation schists which often contain as much as 90 per cent ripidolite. Micro-textures in these schists clearly show the replacement of the original sedimentary material by chlorite. Away from the thrust, the chlorite schists grade quite sharply into sediments which are unaltered except for quartz veining, silicification, and ferruginization, the iron and silica responsible for these phenomena being derived from original sedimentary material displaced by the influx of iron and magnesia.

The chlorite permeation rocks crop out not far from the major thrust mentioned above, forming an almost continuous basic front which has been traced for a distance of over 50 miles parallel to the thrust. It is significant that wherever amphibolites and sediments are involved, similar chlorite schists occur in association with the other major thrusts in Ukinga. In cases where granitic gneisses are thrust over sediments, however, the chlorite schists are completely absent or very poorly developed.

Although the potash metasomatism of Ukinga is not yet proved to be a precursor of granitization, I feel that further work in adjacent areas may show this to be the case. In this part of Tanganyika, therefore, the " exceptional and rare " conditions which, according to Joplin, are required for the production of a regional basic front, are fulfilled, the main factor being the juxtaposition of amphibolites and sediments over a distance exceeding 50 miles. It is hoped that further work to the south will enable this basic front to be traced for an even greater distance.

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10th June, 1952.

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(This communication is submitted with the permission of the Director, Department of Geological Survey, Tanganyika Territory.)

SUPPOSED TUFA BANDS IN CARBONIFEROUS REEF LIMESTONE

SIR,—In a recent paper (*Geol. Mag.*, 1952, lxxxix, 195), Dr. W. W. Black concludes that certain fibrous bands in Carboniferous reef limestone hitherto regarded as primary tufa are in fact secondary crystallization structures in an original calcite mudstone. Dr. Black may well be right in his interpretation of the particular features that he describes in detail, but the varied structures to which the name " reef tufa " have been collectively applied are often not associated with calcite mudstone and some of them are obvious encrustations, whether recrystallized or not.

In a series of papers I have frequently referred to these structures but have never critically discussed their origin, and I have hitherto accepted in a broad way Tiddeman's explanation for the reason that neither I nor others working on reef limestone have found a better one. I have always suspected, however, that the bands are not all of similar origin and that the term " reef tufa " might be inapplicable to some of them.

As Dr. Black observes, the fibrous bands are about $\frac{1}{4}$ in. thick, but this

tends towards a maximum and a single specimen might consist of many parallel or concentric layers varying in depth to less than ½ mm. This is particularly noticeable in those varieties that encrust fossils or that display contortions and other irregularities without obvious encrustation. As regards the extent of the " tufa " I have seen masses several square feet in area on weathered surfaces, and I have observed large colonies of fasciculate lithostrotionids with the corallites cemented together by radiating fibrous calcite in concentric layers and with dark blue-grey calcite mudstone as a final infilling between the encrusted corallites where these are not in actual contact. Such features seem to me to be original, but if any recrystallization has occurred it has been confined to the encrusting material, whether organic or inorganic ; the calcite mud when present is unaffected. If such encrustations were organisms their nature has not been determined. In some cases, as Black points out, bryozoans are associated with the bands and certain of them may in fact be partly recrystallized bryozoa. In this respect it is interesting to note that Dr. Black finds no evidence in support of Dr. J. E. Prentice's statement (*Quart. Journ. Geol. Soc.*, 1951, cvi, 171) that the fibrous bands represent " in recrystallized state the layered skeletons of the reef-forming calcareous algae ". In view of the many references to these structures, notably by Hudson, Bond, and Parkinson, it is rather curious that Prentice does not refer to the prevailing view of their origin.

I do not follow Black in his argument that the formation of tufa necessitates extensive uplifts, since in reef sedimentation much of the growth is near sea level. (See also Bond, *Quart. Journ. Geol. Soc.*, 1950, cv, 157, and *Geol. Mag.*, 1950, lxxxvii, 267.)

Dr. Black's strongest evidence for the secondary nature of the layers is his discovery of fossils partly in calcite mudstone and partly in the fibrous material, which brings me back to my original point of the possible mixed origin of the bands. But whatever the truth of the matter the structures have one thing in common ; they occur only in deposits to which the name reef limestone has been applied. This indicates that the layers are indicative of something inherent in the mode of accumulation of the sediments.

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FOSSILIFEROUS STRATA AT KAPP SCANIA, DAUDMANNSØYRA, VEST SPITSBERGEN

SIR,—During the Birmingham University Expedition to Spitsbergen, 1948 (leader M. F. W. Holland), which was engaged in the geological survey of the southern part of Oscar II Land, B. H. Baker collected some fossils, mainly corals, at Kapp Scania. This locality lies well within a belt extending along the west coast of Spitsbergen which has long been known to consist of practically unfossiliferous folded rocks referred to the Hecla Hoek system of Pre-Devonian age. (A. K. Orvin, Skrifter om Svalbard og Ishavet No. 78, 1940.)

Last year a second collection was made from the same area by a party (including Holland) from the Oxford and Cambridge Spitsbergen Expedition, 1951.

Both collections have now been examined at Cambridge by one of us (C. L. Forbes) and the corals are found to belong to *Caninia calophylloides* (Holtedahl) and related species, while the brachiopods present are probably *Choristites* sp. (the specimens are of poor quality). There are also a few fusulinids, completely silicified, but showing traces of deeply folded septa. The fauna seems to be of high Middle or low Upper Carboniferous age. A comparison of this fauna is being made with the extensive collection of the Cambridge Spitsbergen Expedition, 1949, from the Billefjorden region at present being examined (by C. L. Forbes).

The fossiliferous series has highly metamorphosed rocks of pre-Devonian appearance both to the East and West. Baker, Weiss, and others of the Birmingham Expeditions of 1948 and 1951 have been working on the tectonic implications in conjunction with their work in the St. Jonsfjorden area further north.

MINES AND GEOLOGICAL DEPT., P.O. BOX 339, NAIROBI. — B. H. BAKER.
SEDGWICK MUSEUM, CAMBRIDGE. — C. L. FORBES.
ST. PETER'S HALL, OXFORD. — M. F. W. HOLLAND.

THE AGE OF THE MALVERN FOLDING

SIR,—Perhaps the provider of part of the stimulant (or irritant according to viewpoint) which encouraged Mr. W. Mykura to produce his note on "The Age of the Malvern Folding" (*Geol. Mag.*, Nov.–Dec., 1951), may be allowed to welcome the new facts now available. Even a suggestion that Middle Coal Measures may be represented in the Abberley district is important, because it shows how unwise it is to assume that Middle Coal Measures cannot be present beneath the Trias on the east side of the Malverns.

I think, however, that Mr. Mykura's treatment of the tectonics obscures the unknown. In his text-fig. 2 he boldly shows at Wallhouse a post-Triassic fault with a minimum downthrow of 750 feet to the east, without any supporting evidence. And between Woodbury Hill and Walsgrove Hill he has shown a most remarkable tectonic phenomenon—a *normal* fault which, according to his map (text-fig. 1) follows the bedding for one mile, parallel to the outcrops to the west, turning a 45° corner in the process. Even if such a fault were probable, it requires the easterly outcrop of Aymestry Limestone to be inverted, which should be readily demonstrable, and also the presence of Lower Ludlow on the east. No evidence is given to support either of these requirements, and I am not aware that any has been published. Without such evidence, Groom's original interpretation of the Silurian structure is preferable.

There appears to be a misprint in the middle of p. 390; the dip of 73° to E. 10° S. in the Coal Seam on Ladywood Common, the most important piece of tectonic evidence in favour of post Morganian folding, is shown on the map directed W. 10° S. This dip is placed immediately east of a fault downthrowing 300 feet to the east. Is this throw firmly based, or has it been introduced to allow sufficient room for the steep dips recorded in the Coal Measures to flatten out to the West?

I have often wondered whether some of the folding on the west side of the Malvern line may not be of post-Triassic age. Professor W. F. Whittard (*Geol. Mag.*, 1949, p. 375 and fig. 1) has described post-Triassic thrust fault towards the north-west in the Chepstow area, and Dr. J. S. Turner (*Proc. Geol. Assoc.*, 1927, p. 372) has published evidence for a post-Permian thrust in the Westmorland Pennines. It seems therefore possible that some of the complications with north–south and north-east–south-west strikes in the intervening area may have been produced by post Triassic compressive movements.

N. L. FALCON.

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